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Modeling postbreakup landscape development and denudational history across the southeast African (Drakensberg Escarpment) margin

Peter van der Beek,¹ Michael A. Summerfield,² Jean Braun,³ Roderick W. Brown,⁴ and Alastair Fleming²

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[1] We employ a numerical surface processes model to study the controls on postbreakup landscape development and denudational history of the southeast African margin. Apatite fission track data, presented in the companion paper, suggest that the Drakensberg Escarpment formed by rapid postbreakup river incision seaward of a preexisting drainage divide, located close to its present position, and subsequently retreated at rates of only $\sim 100 \text{ m m.y.}^{-1}$. Numerical modeling results support such a scenario and show that the prebreakup topography of the margin has exerted a fundamental control on subsequent margin evolution. The rheology of the lithosphere, lithological variations in the eroding upper crust, and inland base level falls provided secondary controls. A relatively low flexural rigidity of the lithosphere ($T_e \approx 10 \text{ km}$) is required to explain the observed pattern of denudation as well as the observed geological structure of the southeast African margin. Lithological variations have contributed to the formation of flat-topped ridges buttressing the main escarpment, as well as major fluvial knickpoints. Both these features have previously been interpreted as supporting significant Cenozoic uplift of the margin. An inland base level fall, possibly related to back-cutting of the Orange River drainage system and occurring 40–50 m.y. after breakup, explains the observed denudation inland of the escarpment as well as the development of inland drainage parallel to the escarpment. Our model results suggest that in contrast to widely accepted inferences from classical geomorphic studies, the southeast African margin has remained tectonically stable since breakup and escarpment retreat has been minimal ($< 25 \text{ km}$). **INDEX TERMS:** 1824

Hydrology: Geomorphology (1625); 3210 Mathematical Geophysics: Modeling; 8110 Tectonophysics: Continental tectonics—general (0905); 9305 Information Related to Geographic Region: Africa; **KEYWORDS:** surface process models, landscape development, denudation chronology, Drakensberg Escarpment, Great Escarpment, passive margins

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1. Introduction

[2] Great escarpments along high-elevation passive continental margins are some of the most prominent morphological features on Earth. During the past decade, the importance of understanding the factors controlling escarpment evolution, in order to comprehend better the dynamics of passive continental margins, has become increasingly

appreciated by geologists and geophysicists [e.g., Beaumont *et al.*, 2000; Gilchrist and Summerfield, 1990, 1994; van der Beek *et al.*, 1995]. At the same time, the geomorphological community has shown a renewed interest in large-scale, long-term landscape development of passive margins and other intraplate settings [e.g., Summerfield, 2000]. This has led to the realization that the geomorphic evolution of escarpment systems, which was traditionally interpreted in terms of pulses of uplift and escarpment retreat [King, 1962; Ollier, 1985] may be considerably more complex and variable [Brown *et al.*, 2000; Gallagher and Brown, 1997; Gilchrist and Summerfield, 1994].

[3] An additional impetus for quantifying spatiotemporal patterns of denudation and landscape development has arisen from recent attempts to explain the high topography of southern Africa, amongst other areas, in terms of shallow or deep mantle processes [Gurnis *et al.*, 2000; Lithgow-Bertelloni and Silver, 1998; Nyblade and Robinson, 1994].

¹Laboratoire de Géodynamique des Chaînes Alpines, Université Joseph Fourier, Grenoble, France.

²Department of Geography, University of Edinburgh, Edinburgh, UK.

³Research School of Earth Sciences, Australian National University, Canberra, Australia.

⁴School of Earth Sciences, University of Melbourne, Melbourne, Victoria, Australia.

The constraining of such models based on inferred changes in elevation of the land surface clearly requires that the evidence for such landscape change be securely founded.

[4] Traditionally, the long-term denudational histories of passive margins and adjacent continental interiors have been inferred by linking offshore stratigraphic sequence boundaries to onshore remnants of erosion surfaces, constrained by rare onshore occurrences of dated sedimentary deposits [e.g., King, 1962; Ollier and Pain, 1997]. This approach has been expanded by including the analysis of weathering deposits and duricrusts to characterize erosion surfaces interpreted to be of a particular age [e.g., Gunnell, 1998; Partridge and Maud, 1987; Widdowson, 1997]. A serious problem with these traditional methods is their inherent lack of dating control and therefore the strong reliance on correlation of surfaces based on morphological or sedimentological characteristics, although progress has recently been made in the isotopic dating of weathering deposits [e.g., Vasconcelos, 1999].

[5] Over the past decade, the development of apatite fission track thermochronology [e.g., Bohannon et al., 1989; Brown et al., 1990; Gallagher et al., 1994b; Menzies et al., 1997] and, more recently, cosmogenic isotope analysis [Bierman and Caffee, 2001; Cockburn et al., 2000; Fleming et al., 1999; van der Wateren and Dunai, 2001], has contributed significantly to quantifying the denudational history of passive margins. At the same time, our understanding of what factors control landscape evolution has benefited from the development of numerical models that aim to link lithospheric processes (tectonic uplift and subsidence, flexural isostasy), to those operating on the surface of the Earth (erosion, transport and deposition of sediments) (see Beaumont et al. [2000] for a review). Early, two-dimensional versions of such models have demonstrated the importance of denudation and the resulting isostatic rebound in generating, maintaining, and modifying the morphology of passive margin upwarps [Gilchrist and Summerfield, 1990; ten Brink and Stern, 1992; van der Beek et al., 1995]. More sophisticated, planform, multi-process models have been employed to investigate both the conditions that are necessary to generate and maintain escarpments on high-elevation passive margins [Kooi and Beaumont, 1994; Tucker and Slingerland, 1994], as well as the controls exerted by factors such as lithology and prebreakup morphology on the subsequent evolution of such margins [Gilchrist et al., 1994; Kooi and Beaumont, 1994, 1996; van der Beek and Braun, 1999].

[6] Here, we set out to establish the controls on post-breakup landscape development and the denudational history of the southeast African (Drakensberg Escarpment) margin, using a numerical surface processes model. In the companion paper by Brown et al. [2002], the spatial and temporal patterns of postbreakup denudation along a transect across this margin have been documented using apatite fission track data derived from both surface and deep borehole samples. These data, together with recent estimates of short-term rates of down-wearing and escarpment retreat from cosmogenic isotope analysis [Fleming et al., 1999, Fleming, 2000], are incompatible with traditional views on the evolution of the Drakensberg Escarpment. According to these "classic" views, the Drakensberg Escarpment evolved through parallel retreat from an initial location at the coast-

line [King, 1962; Ollier and Marker, 1985; Partridge and Maud, 1987]. Brown et al. [2002] propose an alternative model of landscape development, in which an escarpment was initiated at the coast but was then rapidly destroyed by rivers flowing from an interior drainage divide. This divide would have existed at a local high on the Karoo basalt plateau just seaward of the present-day Drakensberg Escarpment. Here we evaluate this hypothesis using our numerical model; we also assess the influence of possible secondary controls, such as lithological variation, the flexural rigidity of the lithosphere and the evolution of the inland base level, on postbreakup landscape development across the Drakensberg Escarpment margin.

[7] We first describe the study area and review the existing morphological, stratigraphic, and thermochronological evidence for its denudational and landscape evolution history. From these observations, we distil a conceptual model that serves as a framework for our numerical model simulations. We then introduce the numerical model and present our results. Finally, we discuss the insights that this model provides into the tectonic and geomorphic history of the southeast African margin.

2. Study Area

2.1. Morphology

[8] The sheared margin of southeast Africa (Figure 1) was formed by opening of the Natal Basin and southwestward movement of the Falkland Plateau along the Agulhas Fracture Zone about 130 Ma [Ben-Avraham et al., 1993; Martin and Hartnady, 1986]. The morphology of southeast Africa is characteristic of a high-elevation passive margin, with a prominent erosional escarpment (the Drakensberg Escarpment) separating the highstanding continental interior (the Lesotho Highlands) from a strongly dissected coastal region. Mean elevation rises progressively from the coastline to around 1600 m at ~120 km inland and then more gradually up to around 2000 m at the base of the Drakensberg Escarpment, ~150 km inland (Figures 1 and 2). The escarpment summit, which largely coincides with the continental drainage divide of southern Africa, reaches elevations of around 2700 m in the region of our transect, but rises up to nearly 3500 m to the north of the study area. Mean elevations remain high (around 2700 m) throughout the Lesotho Highlands, up to 300 km inland, before dropping off very gradually to ~1000 m in the continental interior. Mean local relief, defined as the maximum elevation difference within $10' \times 10'$ areas [Summerfield, 1991] is >1000 m throughout the coastal region and the Lesotho Highlands, but drops off sharply to values <500 m in the continental interior of southern Africa.

[9] Relatively linear river systems, which have their headwaters in the escarpment region, drain the area seaward of the continental divide. The main drainage system of the Lesotho Highlands is constituted by the headwater tributaries of the Orange River (see Figure 1) which predominantly flow southwestward, that is, subparallel rather than perpendicular to the escarpment, and deeply incise the highlands.

2.2. Geology and Structure

[10] The geology of southeast Africa is dominated by the late Carboniferous-Triassic Karoo Supergroup (Figure 2)

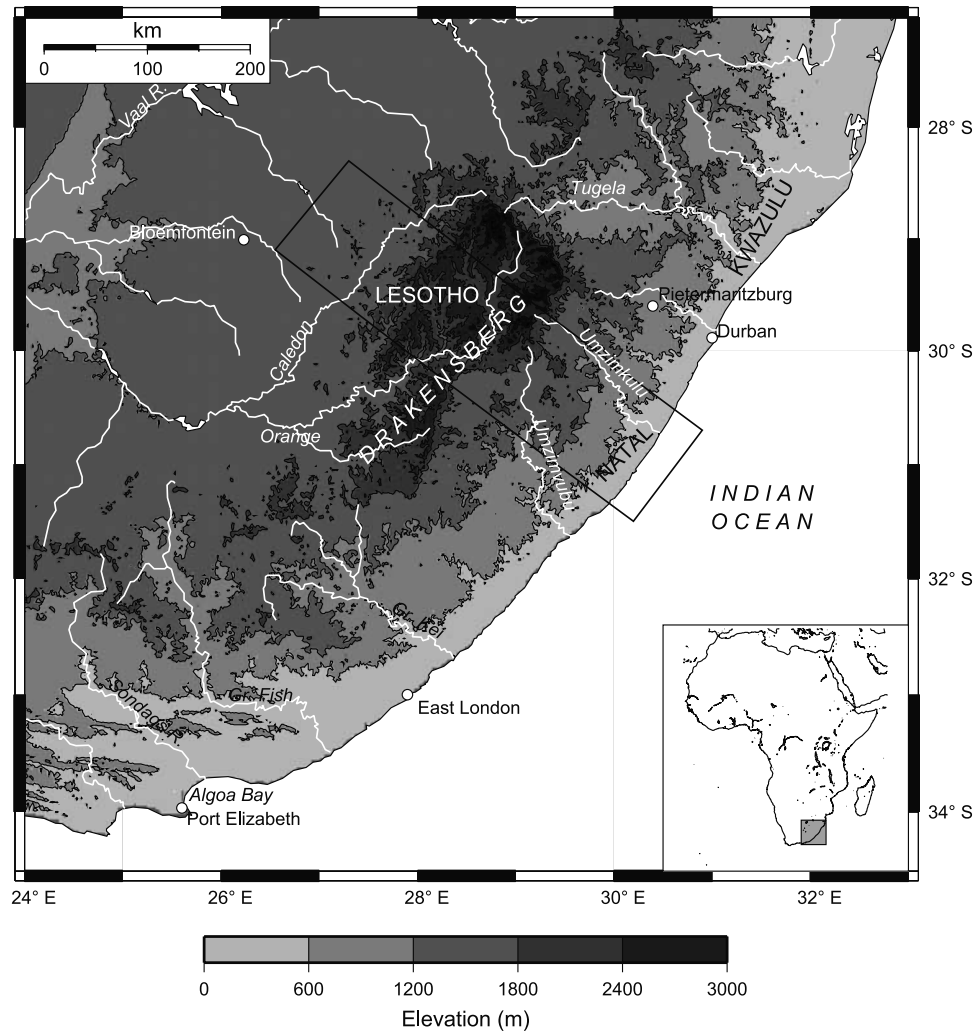


Figure 1. Shaded relief map of the southeastern African margin, color coded according to elevation and showing localities referred to in the text. Box indicates location of transect shown in Figure 2; inset map shows location of study area on the African continent. See color version of this figure at back of this issue.

that overlies Early Paleozoic sediments of the Natal Group and metamorphic basement rocks of the Natal Metamorphic Belt. The Karoo sequence is capped by the extensive Lower Jurassic (183 ± 1 Ma [Duncan *et al.*, 1997]) Drakensberg Basalt, into which the escarpment is cut. The Drakensberg Basalt reaches a maximum thickness of around 1000 m within the study area, but another 500–1000 m has been removed by denudation (see Brown *et al.* [2002] for an evaluation of these numbers). The Karoo Supergroup, the thickness of which reaches ~ 2400 m in the study area, is traditionally subdivided into (1) the Carboniferous Dwyka Formation tillites, (2) shale, siltstone and sandstone of the Permian Ecca and Beaufort Groups, and (3) the Triassic Stormberg Group, consisting of fluvial sandstone, siltstone and shale [cf. Smith, 1990]. Flat-topped ridges protruding from the main Drakensberg Escarpment express variations in resistance between the various lithologies of the upper Karoo strata and extensive Early Jurassic dolerite sills that intrude them. The Karoo sediments are nearly horizontal inland and beneath the Drakensberg Basalt, but are tilted westward across most of the coastal region, exposing

progressively older strata until basement is reached ~ 30 km from the coast. In detail, the coastal region of Natal is characterized by a complex arrangement of arcuate normal and dextral strike-slip faults [Maud, 1961; von Veh and Anderson, 1990], forming a series of horsts and half-graben, in which outliers of Natal and Ecca Group sediments have been preserved.

3. Conceptual Models for Landscape Development in Southeast Africa

3.1. “Classic” Polycyclic Models

[11] The southeast African margin and the Drakensberg Escarpment have played a prominent role in the classic literature on large-scale, long-term landscape development. The geomorphic history of southern Africa has traditionally been interpreted in terms of polycyclic scenarios, following the highly influential works of F. Dixey [e.g., Dixey, 1955] and especially L.C. King [e.g., King, 1951, 1962]. Within our study area, the “Natal Monocline” effectively represents the “type locality” for King’s model of landscape

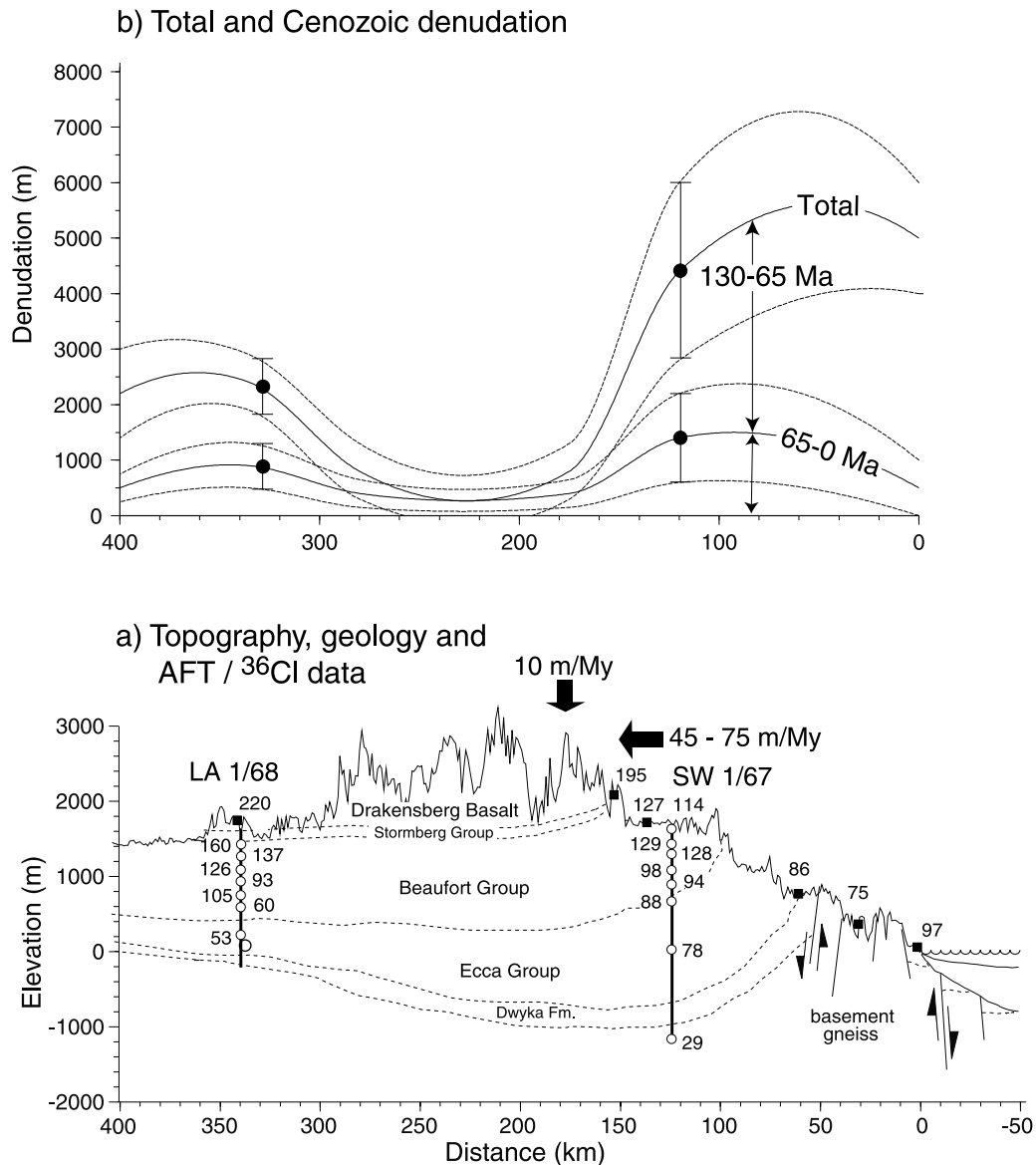


Figure 2. (a) Profile along the transect of Figure 1 showing topography along a central profile (1-km resolution), geology and apatite fission track ages, modified from *Brown et al.* [2002]; arrows with numbers indicate estimated short-term ($\sim 10^4$ years) rates of surface lowering and escarpment retreat from cosmogenic ^{36}Cl data [*Fleming et al.*, 1999; *Fleming*, 2000]. Location of Swartberg (SW 1/67) and Ladybrand (LA 1/68) wells is also indicated. The Dwyka Formation, Eccca, Beaufort, and Stormberg Groups together make up the Karoo Supergroup. (b) Estimated amounts of total (post-130 Ma) and Cenozoic (post-65 Ma) denudation along the transect from modeling of apatite fission track data as well as independent estimates. See *Brown et al.* [2002] for details.

evolution [King, 1982]. In this framework, several erosion surfaces were distinguished, the development of which was supposed to be related to pulses of tectonic uplift and ensuing cycles of erosion. Long-standing controversies have existed concerning the significance, dating and correlation of erosion surfaces, as well as their mode of development, either through plateau downwearing or escarpment retreat (see *Partridge and Maud* [1987] for a review). *Pugh* [1955] and *King* [1955, 1962] proposed that the isostatic response to onshore erosion and offshore deposition is the mechanism responsible for the cycles of landscape evolution. Their work was, however, plagued by apparently

fundamental misconceptions about the nature of isostatic rebound; it was suggested, for instance, that an isostatic response occurred only when a threshold distance of escarpment retreat of ~ 400 km had occurred, rather than progressively in response to denudational unloading [*Gilchrist and Summerfield*, 1991].

[12] The most significant current polycyclic landscape evolution model for southern Africa is that proposed by *Partridge and Maud* [1987] and most recently summarized by *Partridge and Maud* [2000]. In modifying King's original schema, these authors recognize three major erosion surfaces, the "African" surface of Early Cretaceous to Miocene age,

and two “post-African” surfaces initiated in the Miocene (post-African I) and late Pliocene (post-African II), respectively. The sequence of erosion cycles producing the African surface is thought to have begun in response to base level lowering associated with rifting and continental breakup around southern Africa in the Late Jurassic-Early Cretaceous. The African surface is interpreted as having been uplifted by up to 150–300 m in the Early Miocene, an event which promoted the development of the post-African I surface. Partridge and Maud [1987] also argued that both the African and post-African I surfaces were further uplifted in the late Pliocene by up to 900 m. In both the proposed Miocene and late Pliocene events, the axis of maximum surface uplift was considered to be aligned approximately parallel with, and about 80 km west of, the present Indian Ocean coastline.

[13] An alternative scenario has been proposed by Burke [1996], in which the high elevation of southern Africa and the escarpment are considered to have arisen from rapid surface uplift ~ 30 m.y. ago. This model is based on the assumption that Late Cretaceous-Eocene marine deposits encountered on the coastal strip of South Africa [Partridge and Maud, 1987] once covered the entire southern African plateau. Unfortunately, these sediments, if they ever existed, have since been removed by erosion, making Burke’s [1996] model impossible to test with reference to the displacement of marine deposits.

3.2. Criticism of Polycyclic Models

[14] Critics of the landscape evolution models proposed by L.C. King have included Wellington [1955] and De Swardt and Bennet [1974]. Wellington [1955] questioned the paradigm of widespread preservation of erosion surfaces and pointed out the importance of lithological controls on the step-like morphology of large areas of the southern African landscape. De Swardt and Bennet [1974], while remaining within a polycyclic framework, emphasized the duality of inland and coastal landform development in southeast Africa and questioned the possibility of correlating erosion surfaces encountered inland and seaward of the Drakensberg Escarpment.

[15] More recently, Summerfield [1985] [see also Summerfield, 1996] has pointed out serious problems in distinguishing interpretation from observation that are inherent to the recognition, correlation and dating of erosion surfaces. He also argued that polycyclic landscape evolution models are incompatible with the sedimentary record contained in the offshore basins of southern Africa. Apart from the breakup event itself, the timing of the proposed surface uplift events do not appear to correlate with increases in rates of sediment supply to the margins of southern Africa [e.g., Dingle *et al.*, 1983; Rust and Summerfield, 1990; Brown *et al.*, 1990]. However, the interpretation of sediment supply rates, as represented by offshore sediment volumes, is difficult to relate directly to spatiotemporal variations in continental erosion rates because of the possibility of changing source areas through time [Rust and Summerfield, 1990].

3.3. A Revised Model for Landscape Evolution on the Southeast African Margin

[16] The apatite fission track thermochronology (AFT) data presented by Brown *et al.* [2002], in combination with recently acquired cosmogenic isotope data [Fleming *et al.*,

1999; Fleming, 2000], are inconsistent with the modern polycyclic models for the evolution of the Drakensberg Escarpment [e.g., Partridge and Maud, 1987, 2000] for two reasons.

[17] First, although supporting the notion of a major denudational episode initiated by breakup along the southeast African margin, neither the AFT data nor the offshore sedimentary record suggest the large-scale Cenozoic surface uplift events proposed by Partridge and Maud [1987, 2000]. Although AFT data cannot be used directly to quantify surface uplift [Summerfield and Brown, 1998], a significant increase in elevation along the southeast African margin would be expected to have generated a significant denudational response and therefore increased sediment flux [Brown *et al.*, 1994; Kooi and Beaumont, 1996]. Accelerated denudation in the late Cenozoic would be an expected outcome of models calling for late-stage surface uplift along the margin of southeast Africa [e.g., Burke, 1996; Partridge, 1997], since such an event would have significantly increased local relief near the coast. Although the removal of up to a kilometer of overburden would be very difficult to detect using AFT in surface samples, the borehole samples which are currently at temperatures of ~ 40 – 70°C would be extremely sensitive to late Cenozoic accelerated denudation and cooling.

[18] Second, as explained in detail by Brown *et al.* [2002], the AFT data are not consistent with the proposed evolution of the Drakensberg Escarpment through parallel retreat at a constant rate from an initial position close to the coastline. The AFT data from the Swartberg borehole, located ~ 30 km seaward of the escarpment, indicate that a phase of accelerated denudation occurred between 90 and 70 Ma, much older than what would be expected if the escarpment had retreated at a constant rate since its inception during continental breakup. In addition, cosmogenic ^{36}Cl analysis of samples collected from the escarpment free face [Fleming *et al.*, 1999; Fleming, 2000] indicates that short-term ($\leq 10^4$ years) rates of escarpment retreat are in the range of 45 – 75 m.m.y. $^{-1}$, with an absolute maximum of ~ 200 m.m.y. $^{-1}$, an order of magnitude lower than the rates expected in a constant retreat scenario (>1 km.m.y. $^{-1}$).

[19] On the basis of the above arguments and recent quantitative modeling of landscape development on passive margins [Gilchrist *et al.*, 1994; Kooi and Beaumont, 1994, 1996], we propose a revised conceptual model for landscape development on the southeast African margin. We suggest that prior to continental breakup in the Early Cretaceous, the region was at a mean elevation of ~ 2500 m (see discussion of this number below) and has remained relatively tectonically stable since that time. We envisage the existence of a prebreakup drainage divide located close to the present position of the Drakensberg Escarpment, possibly formed by thickness variations in the pile of Drakensberg basalts. Continental breakup resulted in a major fall in base level and rapid destruction of the basalt-capped plateau seaward of the preexisting divide. This is recorded as an early phase of accelerated cooling in the AFT samples seaward of the escarpment [Brown *et al.*, 2002]. We suggest that the present-day escarpment developed at this divide and, once formed, retreated inland at rates of only ~ 100 m.m.y. $^{-1}$. The slightly delayed phase of accelerated denudation inland of the escarpment may be associated with headward extension of the Orange River drainage system, breaching of the

basalt cover by these tributaries and subsequent rapid incision and relief formation.

[20] We can make an estimate of the prebreakup topography of the margin by flexurally backstacking the amount of removed overburden, assuming there has been no subsequent tectonic displacement [cf. *van der Beek et al.*, 1994]. The paleotopography thus calculated [Brown et al., 2002], using the amounts of denudation from the AFT data as well as independent estimates of the depth of denudation of the Drakensberg Basalt, and realistic estimates for the flexural rigidity, is characterized by a drainage divide occurring at ~ 120 km inland and ~ 2750 m elevation, consistent with our conceptual model.

[21] We will use our conceptual model for landscape development in southeast Africa as a starting point for our numerical simulations in the remainder of this paper. Note, however, that our modeling results will not prove our conceptual model “right” or “wrong.” This is because of the inherent non-uniqueness of the modeling results, as well as the inherent inability of the models to capture the full complexity of earth surface processes [e.g., *Oreskes et al.*, 1994]. Our modeling exercise does, however, serve to quantitatively establish (1) to what extent our conceptual model is supported by the available data and (2) what are the important factors that control the morphological evolution and denudational history of the margin.

4. Numerical Model

4.1. Model Outline

[22] We employ the numerical surface processes model Cascade [Braun and Sambridge, 1997; *van der Beek and Braun*, 1998, 1999], which uses a spatial discretization of the evolving landscape as a large number of interconnected irregular cells. Within this model, as in other surface process models (see *Beaumont et al.* [2000] for a review) three types of processes are supposed to control large-scale, long-term landscape development: hillslope regolith transport, bedrock landsliding, and fluvial transport. Regolith transport processes are modeled by a linear diffusion law in which the rate of erosion or deposition is proportional to topographic curvature:

$$\partial h / \partial t = \kappa_D \nabla^2 h, \quad (1)$$

where h is elevation, t is time, and κ_D is a linear diffusion coefficient with unit $\text{m}^2 \text{yr}^{-1}$. We include bedrock landsliding as a second type of slope process that operates when a critical slope S_c is surpassed. The bedrock landsliding algorithm is designed to conserve mass and is explained in detail by *van der Beek and Braun* [1999]. The algorithm is deterministic, which may not be particularly realistic for landsliding, but given the relatively coarse spatiotemporal resolution of our models, we believe this is a reasonable simplification.

[23] Fluvial processes are driven by uniform precipitation, channeled along streams that follow the route of steepest descent from their source to one of the sides of the model. The stream power law dictates the carrying capacity q_f^{eqb} of the streams:

$$q_f^{\text{eqb}} = K_f q_r \partial h / \partial l \quad (2)$$

and is controlled by a dimensionless fluvial transport coefficient K_f , the discharge q_r [$\text{m}^3 \text{yr}^{-1}$] and the local stream gradient $\partial h / \partial l$. Local discharge is calculated by spatially integrating the upstream precipitation v_r for every cell in the model [cf. *Braun and Sambridge*, 1997]. The streams erode or deposit material according to the balance between q_f^{eqb} and the sediment flux q_f resulting from upstream erosion:

$$\partial h / \partial t = (q_f - q_f^{\text{eqb}}) / w L_f, \quad (3)$$

where w is river width (assumed constant) and the erosion length scale L_f is a measure of the “detachability” of the substratum, included to model supply limited behavior. L_f takes different values for bedrock (L_{fb}) and alluvium (L_{fa}); in general, $L_{fb} \gg L_{fa}$.

[24] Given the spatial scale of our models (several thousand km^2), we include the isostatic response to denudation, which is calculated by solving the two-dimensional flexure equation:

$$D \nabla^4 W_{(x,y)} + \rho_m g W_{(x,y)} = -\rho_c g \Delta h_{(x,y)}, \quad (4)$$

where ρ_c and ρ_m are the upper crustal and mantle densities [kg m^{-3}], respectively, g is the acceleration of gravity [m s^{-2}], D is the flexural rigidity of the lithosphere [N m] and W is the flexural deflection [m].

4.2. Parameter Values and Boundary Conditions

[25] We model a 100-km-wide transect across the margin. Our initial models concentrate on the evolution of the region seaward of the escarpment and are 250 km long. Subsequent models, in which we also consider the inland region, have a length of 400 km. The sides perpendicular to the escarpment act as reflective boundaries, whereas both sides parallel to the escarpment are sediment “sinks” (where $\partial h / \partial t = 0$). We assume the coastline to be fixed but allow the inland boundary of the model to move vertically with isostatic rebound. All of the models are run for 130 m.y., approximately the time since breakup [Ben-Avraham et al., 1993]. The initial topography of the margin is constrained from flexural backstacking as discussed in section 3.3.

[26] Values for the erosional parameters K_f , ν_R , κ_D , L_f , and S_c are very difficult to evaluate because they represent the integrated effect of a combination of surface processes. Moreover, the spatial and temporal scales to which these parameters pertain are such that they cannot be directly compared to field measurements of “diffusivity” or fluvial transport capacity [e.g., *Heimsath et al.*, 1997; *Stock and Montgomery*, 1999]. We therefore let these parameters vary within the limits provided by our previous experiences. Parameter values that lead to acceptable model results (Table 1) are all within an order of magnitude of values employed in modeling the southeast Australian margin [van der Beek and Braun, 1998, 1999], which occurs in a similar tectonic and climatic setting. This provides us with confidence in the model formulation that we have adopted.

[27] The amount and wavelength of isostatic rebound are determined by the flexural rigidity of the lithosphere, as expressed by its equivalent elastic thickness (T_e), and have a large influence on the morphology of the margin [Gilchrist

Table 1. Parameter Values and Initial/Boundary Conditions for Numerical Experiments^a

Parameter	Value			Lithological Variation
	Plateau Degradation	Escarpment Retreat	Low T_e	
Initial position of escarpment, km	125	0	125	125
Fluvial transport coefficient times precipitation $K_f v_R$, $\times 10^{-3}$ m yr ⁻¹	2	5	2	2.5
River width times bedrock erosion length scale $w L_{fb}$, km ²	100	100	100	100/20
River width times alluvial length scale $w L_{fa}$, km ²	10	10	10	10
Hillslope diffusion coefficient κ_D , m ² yr ⁻¹	10^{-3}	10^{-3}	10^{-3}	$10^{-3}/10^{-2}$
Threshold slope for landsliding S_c	10°	10°	10°	10°
Equivalent elastic thickness of the lithosphere T_e , km	30	30	10	10

^aParameter values for the inland base level fall model are similar to those for the lithological variation model.

and Summerfield, 1990; Kooi and Beaumont, 1994; van der Beek et al., 1995]. T_e estimates for southeast Africa vary from 72 km from coherence analysis of Bouguer gravity and topography on the Kaapvaal Craton [Doucouré et al., 1996] to ~ 10 km from forward modeling of the gravity signature of the offshore margin [Watts and Marr, 1995]. Hartley et al. [1996] suggest that southeast Africa is characterized by a strong variation in T_e from very high values inland to near-zero values offshore. We assess the influence of elastic thickness by comparing models with $T_e = 30$ km and $T_e = 10$ km. For reasons of numerical efficiency, we restrict ourselves to modeling a uniform-thickness continuous elastic plate, for which isostatic rebound can be calculated using spectral methods [Braun and Sambridge, 1997]. It has been argued that morphological differences between passive margins may be explained by these being underlain by either broken or continuous elastic plates [e.g., Stiwe, 1991; ten Brink and Stern, 1992]. In the case of the Drakensberg Escarpment, however, we expect these differences to be minimal, because of the relatively low flexural rigidities employed as well as the large width (~ 150 km) of the deeply denuded region seaward of the escarpment.

[28] We do not include the effects of tectonic loads induced by shearing nor those of offshore sedimentation [van Balen et al., 1995; van der Beek et al., 1995]. These loads are not well constrained and considered of relatively minor importance. Total sediment thickness in the Natal Basin does not exceed 2500 m [Martin, 1987]. Furthermore, the effect of lateral heat conduction from hot oceanic to colder continental lithosphere at transform margins is felt within the seawardmost 30 km of continental lithosphere [Gadd and Scrutton, 1997], approximately the width of the offshore southeastern African margin.

5. Modeling Results

[29] We have run several tens of different models that vary in their input parameters and boundary conditions. Only those that we believe to have the most direct relevance to landscape development on the southeast African margin are reproduced here (Table 1).

5.1. Initial Position of the Drainage Divide

[30] Our first set of model runs was designed to evaluate the influence of the inferred preexisting drainage divide. In the “plateau degradation” model, the initial topography is that obtained by flexural backstacking of the inferred amounts of denudation [Brown et al., 2002]: the initial elevation rises from 2 km at the coastline to 2.75 km at the

preexisting drainage divide, 125 km inland, before descending to 2.5 km at the inland model boundary. An alternative model (the “escarpment retreat” model) is presented in which the initial elevation consists of a flat, 2.5-km-high plateau. In both models, a base level fall down to sea level is introduced at the coastline at the onset of the model run. These models are very similar to those presented by Gilchrist et al. [1994] and Kooi and Beaumont [1996] and our results confirm their conclusions about landscapes dominated by escarpment retreat (equivalent to our escarpment retreat model) compared with those arising from differential down-wearing (our plateau degradation model). Parameter values for both models are the same, except for the fluvial transport coefficient $K_f v_R$, which was calibrated so that in both models the escarpment is located ~ 150 km inland after 130 m.y.

[31] The evolution of both models is dramatically different (Figures 3 and 4) because of the difference in contributing drainage area (and thus incision power) of the rivers flowing over the initial escarpment. In the plateau degradation model, the area between the divide and the coast is incised rapidly by seaward flowing rivers. This leads to destruction of the initial escarpment within 10 m.y. and the creation of a new escarpment at the locus of the initial drainage divide after ~ 70 m.y. Because river heads are now pinned at the divide, this escarpment retreats inland at a rate of only ~ 300 m m.y.⁻¹ (Figure 5). In the escarpment retreat model, in contrast, erosion is concentrated at the locus of the escarpment throughout the evolution of the model. Consequently, there is nearly steady state retreat of the escarpment, at a rate of ~ 1 km m.y.⁻¹.

[32] Drainage development is also very different for both models. In the plateau degradation model, drainage is immediately set up to flow away from the initial drainage divide, leading to linear river systems draining both inland and seaward. Once the escarpment is established, it retreats by reversal of the headwaters of the inland-draining rivers. The seaward draining rivers preferentially follow the valleys previously incised in the upland area, leading to a jagged but relatively straight escarpment. In the escarpment retreat model, the initially flat area inland of the escarpment has strongly disorganized drainage, which only becomes integrated as flexural isostatic rebound imposes an inland dip on the plateau. This leads to frequent capture of internal drainage areas by the streams flowing seaward and, consequently, much more dendritic drainage patterns and an escarpment with large embayments.

[33] Maximum denudation is slightly higher in the escarpment retreat than in the plateau degradation model (2.6 versus 2.25 km) but in the latter model, a wider area is eroded

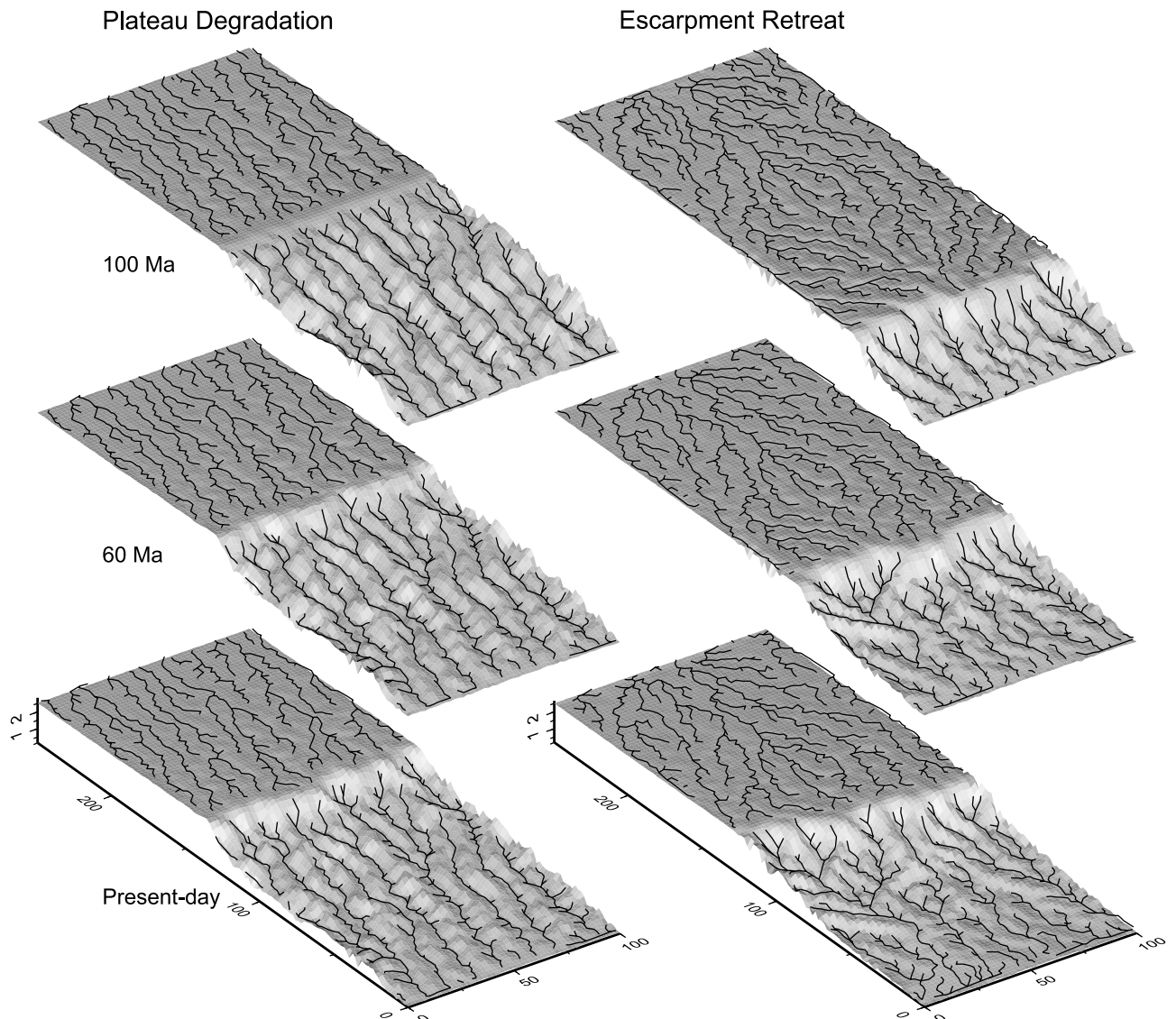


Figure 3. Artificially illuminated oblique views of the topography and drainage patterns predicted by the plateau degradation and escarpment retreat models, at 100 Ma, 60 Ma, and the present-day. Axes indicate model coordinates in kilometers. Parameter values for these models are given in Table 1. See color version of this figure at back of this issue.

uniformly. A comparison of the denudation histories for a point 30 km seaward of the escarpment (Figure 6) confirms our qualitative expectation: total denudation at this point is similar (1.8 km) for both models, but the onset of denudation occurs much later (50 Ma) in the escarpment retreat model than in the plateau degradation model (110 Ma). Whereas denudation rates at this point are nearly constant through time in the plateau degradation model, a clear pulse of denudation can be correlated with the passage of the escarpment in the escarpment retreat model. As expected, the plateau degradation model fits the observations better than the escarpment retreat model, and we will retain the initial conditions corresponding to this model in our following model runs.

5.2. Flexural Rigidity

[34] In both the above models, the predicted amount of denudation seaward of the escarpment is significantly less

than that observed. The models predict a maximum depth of denudation on the coastal strip of ~ 2.5 km, whereas the fission track data of *Brown et al.* [2002] constrain this to be >4 km and the regional geology >3.5 km. The comparison of predicted denudational histories with that inferred from the Swartberg well data (Figure 6) also shows that the models underpredict the amount of denudation.

[35] Sensitivity tests show that the maximum amount of denudation predicted by the models is much less dependent on variations in the erosional parameters $K_f v_R$, κ_D and L_f than the escarpment retreat rate, which is reasonably well predicted by the plateau degradation model. Therefore, varying the erosional parameters will not provide more satisfactory model results. A possible solution for matching the modeled and observed amounts of denudation is that the flexural rigidity of the lithosphere is much lower than that adopted previously.

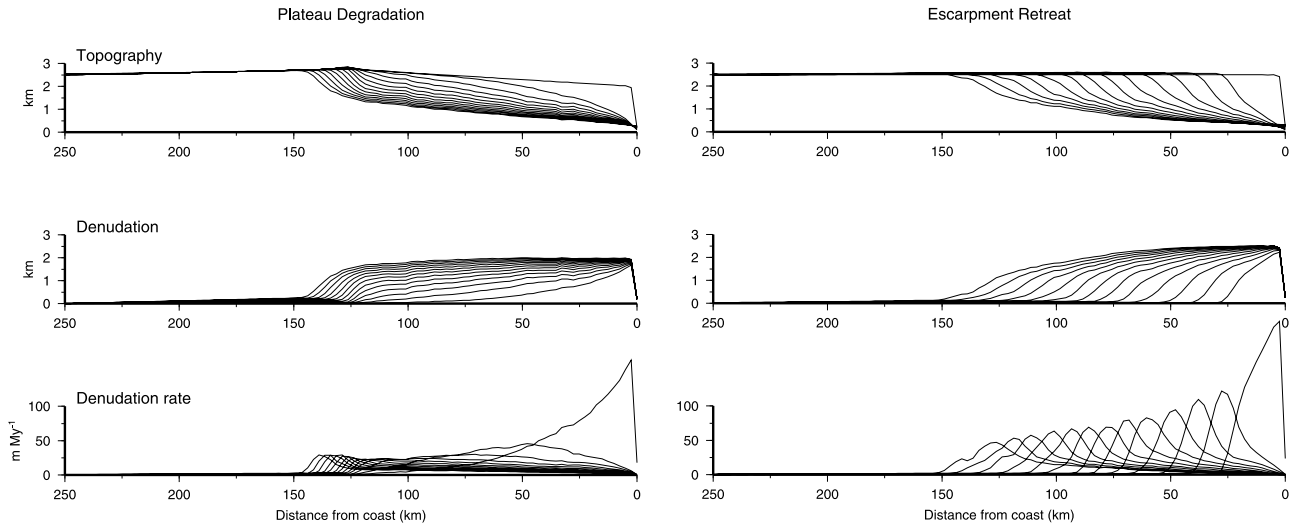


Figure 4. Evolution of strike-averaged topography, denudation, and denudation rates for the plateau degradation and escarpment retreat models. Profiles are shown at 10 m.y. intervals, from zero (start of model run) to 130 m.y. (present-day).

[36] Figure 7 shows the evolution of a model for which all parameters and boundary conditions are similar to the plateau degradation model, except that T_e has been lowered to 10 km. This “low T_e ” model predicts maximum amounts of denudation of 2.75 km, whereas the maximum amount of isostatic rebound is tripled as compared to the original model (1.2 km versus 0.4 km). The wavelength (~ 150 km) and amplitude (1.2 km) of isostatic rebound correspond favorably to the observed westward tilting of Karoo sediments and basement uplift seaward of the escarpment (e.g., Figure 2). The predicted denudation, however, is still not sufficient to expose basement.

[37] The only unambiguous marker of postbreakup surface uplift along the margin is provided by remnants of Late Cretaceous-Eocene marine deposits [Maud and Botha, 2000; Partridge and Maud, 1987] that occur between Port

Elizabeth and East London, ~ 400 km to the southwest of our transect (see Figure 1). Nevertheless, we have evaluated the possibility of such deposits having existed but having subsequently been removed through erosion by comparing the predicted amounts of isostatic rebound over the last 50 m.y. of the model runs to the elevations of these deposits (Figure 8). We subtract a 200 m eustatic sea level fall from

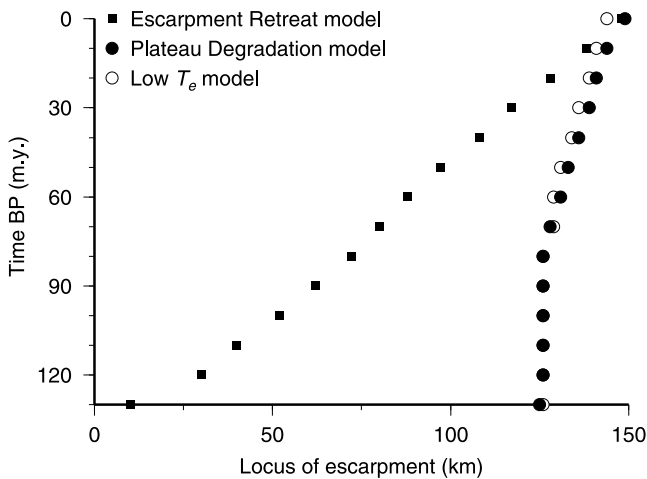


Figure 5. Comparison of escarpment retreat rates for the plateau degradation, escarpment retreat, and low T_e models. The distance from the shoreline to the escarpment is shown as a function of time before present.

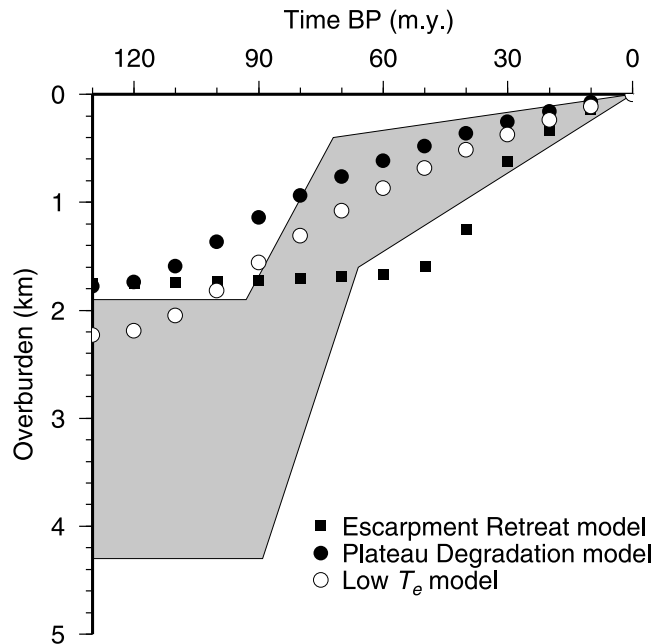


Figure 6. Denudation history for a point located 30 km seaward of the escarpment for the models shown in Figure 5. Plot shows the (strike-averaged) overburden for points now at the surface as a function of time before present. Shaded polygon represents spectrum of acceptable denudation histories from the Swartberg well fission track data [cf. Brown *et al.*, 2002] for comparison with model predictions.

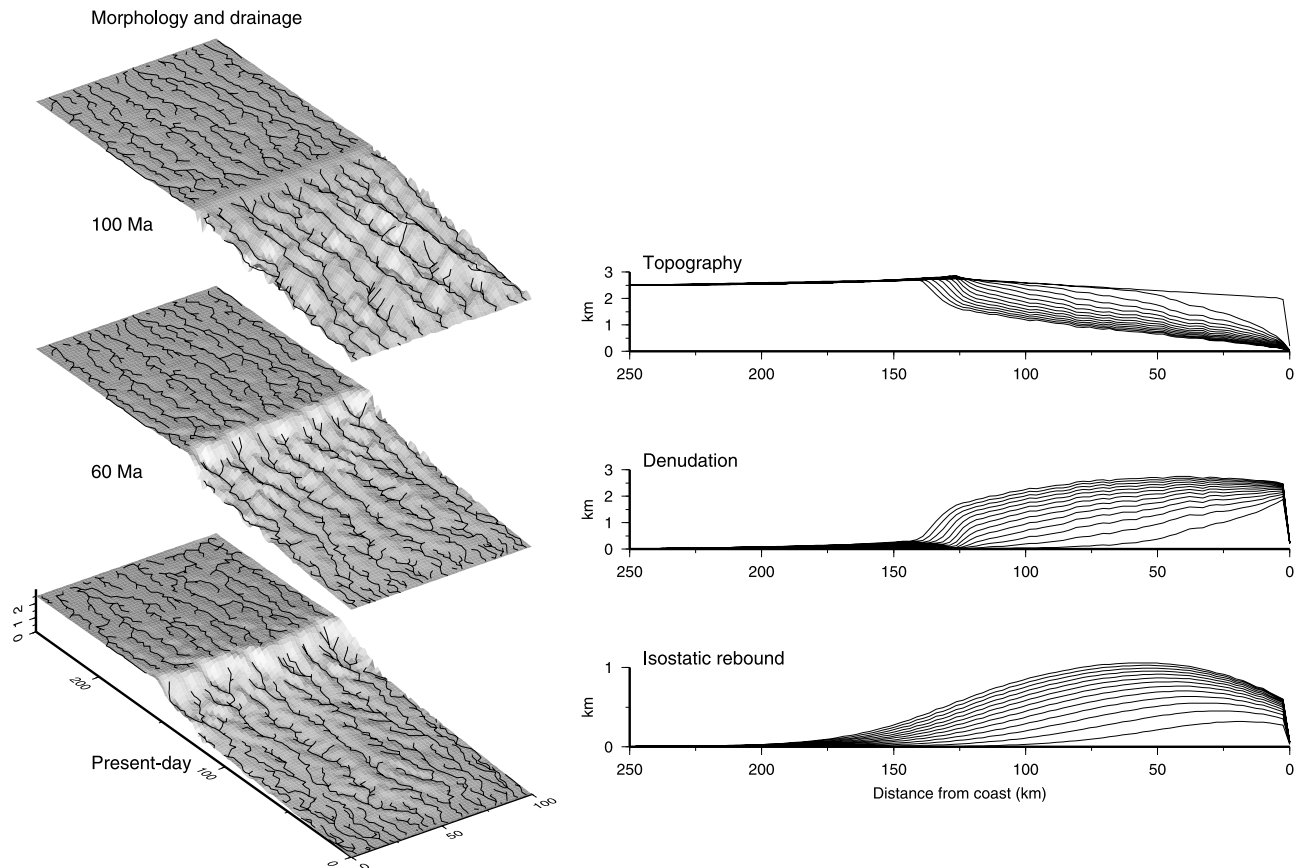


Figure 7. Predicted evolution of low T_e model: (left) artificially illuminated oblique views of the topography and drainage patterns at 100 Ma, 60 Ma, and the present-day; (right) plots of the evolution of strike-averaged topography, denudation, and isostatic rebound at 10 m.y. intervals. Parameters for this model are similar to those of the plateau degradation model, except that T_e has been lowered to 10 km (see Table 1).

the latter, as this provides a mean estimate for eustatic sea level during the Eocene and a lower bound for the Late Cretaceous [Pitman, 1978; Haq *et al.*, 1987]. Our model explains ~80% of the uplift of the highest deposits, but the wavelength of isostatic rebound in our model is much larger than that apparent from the elevations of the marine deposits. This may partly reflect the fact that our model does not include offshore sediment loading, but also suggests that these deposits record local postbreakup vertical motions that have affected the Algoa Basin region. Given that the elevations of all except the most inland of these deposits are actually lower than what we would expect them to be, given a reasonable eustatic sea level change and our predictions for regional isostatic rebound (Figure 8), we would argue that these deposits record local relative subsidence of the Algoa Basin rather than regional uplift.

[38] Although the predicted amount of denudation is still less than observed, the predicted denudational history for this model fits the Swartberg data for times <90 Ma (Figure 6). The larger amount of isostatic rebound in this model also tends to slow down the rate of escarpment retreat to <200 m/m.y. during the last 60 m.y. of the model run (Figure 5). This is because inland gradients at the escarpment are increased, so that inland-flowing rivers can compete more efficiently with those flowing seaward. The predictions of

the low T_e model thus fit the observations significantly better than the previous models.

5.3. Lithological Variation

[39] As discussed previously, lithological variation appears to have influenced the morphology of the southeast African margin. In order to assess the effect of lithological control on the morphological evolution, we have constructed a model in which a less resistant layer, representing the Karoo sediments, lies between two more resistant layers, representing the Drakensberg basalts and the basement, respectively.

[40] The erodibility of a lithological unit is expressed in the model by the erosion length scale L_{fb} . We decreased L_{fb} by a factor of 5 for the Karoo sediments, as compared to the subjacent and superjacent layers (see Table 1). We also enhanced the susceptibility of this unit to slope processes by increasing the diffusion coefficient κ_D by a factor of 10. We take the base of the Drakensberg Basalt to have been initially at 1.5-km elevation, so that the initial thickness of the basalt varies between 0.5 km at the coastline and 1.25 km at the drainage divide. The initial thickness of the Karoo sediments is taken as 2.4 km. We also include a single 200 m thick “resistant” layer embedded within the Karoo sediments (characterized by erosional parameters similar to those

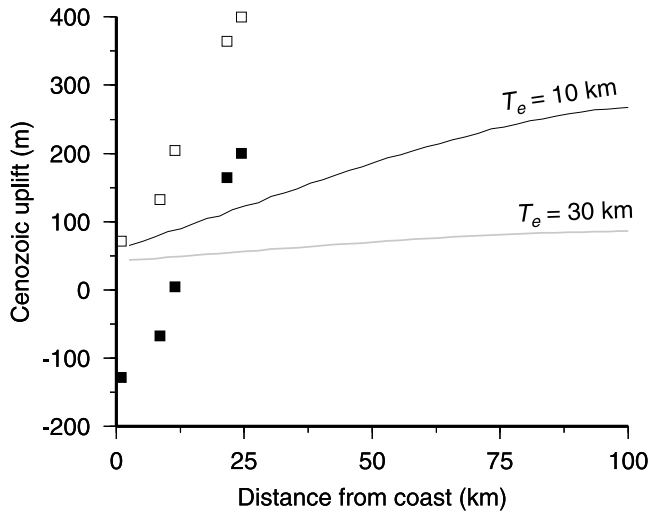


Figure 8. Isostatic rebound as a function of distance from the shoreline integrated over the last 50 m.y. of the model runs for the plateau degradation model (shaded line annotated $T_e = 30$ km) and the low T_e model (solid line annotated $T_e = 10$ km). These model predictions are compared to the elevations of Late Cretaceous-Eocene marine deposits encountered on the coastal strip of south-eastern South Africa [Partridge and Maud, 1987]. Open symbols are the present-day elevations of these deposits; solid symbols include a 200-m Cenozoic eustatic sea level fall subtracted from the present-day elevations.

of the basalt and basement), in order to study the effect of smaller-scale lithological variation within the Karoo sequence and the dolerites intruding it.

[41] Results for this “lithological variation” model are shown in Figure 9. The overall evolution of this model is similar to the (uniform lithology) low T_e model, although total denudation is slightly higher (up to 3.2 km) because of the more easily eroded middle (“Karoo”) layer. The model predicts basement to be exposed between 20 and 50 km inland. The most conspicuous morphological difference is the complete erosion of the finger-like escarpment buttresses that are preserved in the uniform lithology models. Instead, a 5–10 km wide step in the topography evolves from ~90 m.y. onward, which results from the exposure of the resistant layer within the Karoo sedimentary sequence. This topographic step can be recognized in the predicted present-day morphology as narrow, flat-topped ridges buttressing the main escarpment at an elevation of ~1000 m.

[42] Significantly, the model also predicts that lithological variation influences the fluvial long profiles. Figure 10 shows a comparison of a typical river long profile predicted by the lithological variation model with that predicted by the uniform lithology low T_e model, as well as with the observed profile of the Umzimkulu River (one of the rivers that drain from the Drakensberg Escarpment to the Indian Ocean coast). Whereas the uniform lithology models predict the establishment of “graded” fluvial profiles, with a monotonous decrease in river slope downstream, the lithological variation model predicts the development of a ~200-m-high knickpoint where the river crosses the resistant layer within the Karoo sequence. The latter profile

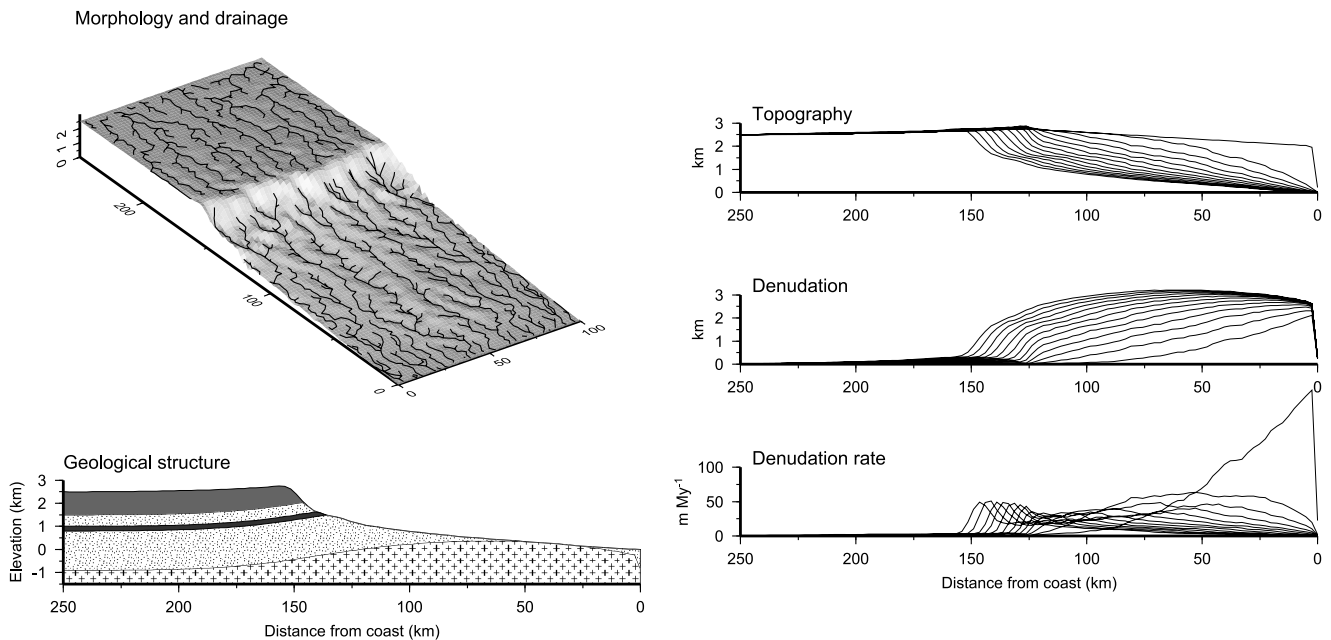


Figure 9. Predicted evolution of the lithological variation model, characterized by 2.4 km of more erodible (Karoo) sediments (dotted) between more resistant top and bottom layers representing the Drakensberg basalt (shaded) and basement (crosses), respectively, and including a 200-m-thick resistant layer within the Karoo sediments (black). (left) Artificially illuminated oblique view of predicted present-day topography and drainage as well as predicted strike-averaged topography and geological structure. (right) Evolution of strike-averaged topography, denudation, and denudation rate.

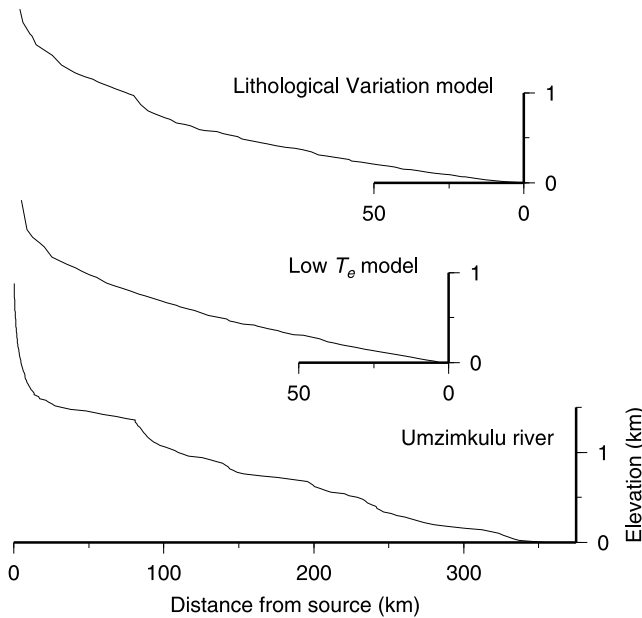


Figure 10. Comparison of characteristic river profiles predicted by the lithological variation and low T_e (uniform lithology) models with the Umzimkulu River of eastern South Africa. The river shown for the lithological variation model has its source at $x = 48$ km, $y = 148$ km; the one for the low T_e (uniform lithology) model has its source at $x = 47$ km, $y = 135$ km. Note change in horizontal scale between predicted and observed river profiles; the observed Umzimkulu river profile is much longer than the modeled profiles because it was digitized at a much higher resolution than the model grid node spacing.

resembles the observed profile of the Umzimkulu River, which is characterized by a series of knickpoints. The most prominent of these occurs where the river crosses a thick Lower Jurassic dolerite [Fleming, 1997].

5.4. Inland Base Level Lowering

[43] So far, we have only been concerned with the evolution of the area seaward of the escarpment. Inland, denudation rates remained close to zero. However, there is evidence that 0.5–1.0 km of overburden has been removed from the top of the Lesotho highlands [Brown *et al.*, 2002]. The denudation of the Lesotho Highlands may have been triggered by lowering of the inland base level due to back-cutting of the westward draining Orange River system. In our final set of models, we study the effect of such inland base level lowering on the evolution of the escarpment. We incorporate an inland base level fall by forcing one of the inland corners of our model (the length of which has been increased to 400 km) to decrease in elevation by 1 km between 85 and 80 Ma. This model also includes lithological variation; in order to reduce complexity, however, the resistant layer within the Karoo sedimentary sequence was not incorporated.

[44] The evolution of the seaward part of the model is very similar to that described before (Figure 11). Inland, the forced drop in elevation of one model corner drives incision of a nearly one km-deep gorge close to the edge of the model. This gorge system captures adjacent drainage

through headward erosion of tributaries and breaching of interfluvies, leading to the progressive collection of inland drainage in one major river system that flows subparallel to the escarpment.

[45] The predicted mean inland denudation for this model is ~ 0.5 km, but this is an average between gorge incision that reaches 1 km and negligible interfluvial lowering. Figure 12 shows a comparison of predicted and observed strike-averaged topography, and predicted present-day geological structure. The predicted topography fits that observed along the transect, although the model predicts a strongly concave topography seaward of the escarpment, whereas a relatively linear rise in elevation is observed. A linear elevation profile between the escarpment and the coastline is promoted by low flexural rigidities [van der Beek *et al.*, 1995] and/or a low fluvial incision length scale [van der Beek and Braun, 1998], suggesting that we are employing maximum estimates for these two parameter values. Lowering either one of them would increase the amount of total denudation. The conspicuous steps in the morphology seaward of the escarpment are not reproduced by the model because it does not include small-scale lithological variation within the Karoo sequence.

[46] Figure 12 also compares the predicted denudation histories with the fission track record from the Swartberg and Ladybrand wells. Predicted denudation histories fit the data for both wells, although the predicted denudation is close to the minimum amount of denudation permitted by the data, and denudation rates between 90 and 60 Ma in the Swartberg well appear to be underestimated. The predicted margin structure, in which a thin layer of basalts persists up to 400 km inland, whereas the observed inland limit of the Drakensberg Basalt lies 300–350 km inland (Figure 2), also suggests that predicted amounts of total denudation are relatively low compared to those observed.

6. Discussion

[47] Our numerical model predictions, when compared to the available empirical data, generally support the conceptual model for the evolution of the southeast African margin outlined here and by Brown *et al.* [2002]. Two key elements of this model are that (1) the escarpment has experienced only limited retreat since breakup and was therefore initiated close to its present-day position rather than at the shoreline or the shelf edge and (2) no large-scale post-breakup tectonic surface uplift is required to explain the morphology and denudation history of the southeast African margin.

6.1. Rate of Escarpment Retreat

[48] A slow escarpment retreat rate, in the order of 100 m m.y.^{-1} rather than 1000 m m.y.^{-1} , is required by the cosmogenic isotope data of Fleming *et al.* [1999] and Fleming [2000]. These data are consistent with the apatite fission track data from the Swartberg well, 30 km seaward of the escarpment, which indicate a phase of accelerated denudation at around 90–70 Ma, much earlier than expected if the escarpment retreated from an original position close to the shoreline to its present-day location.

[49] Our models show how, if the prebreakup topography of the margin included a drainage divide located close to the

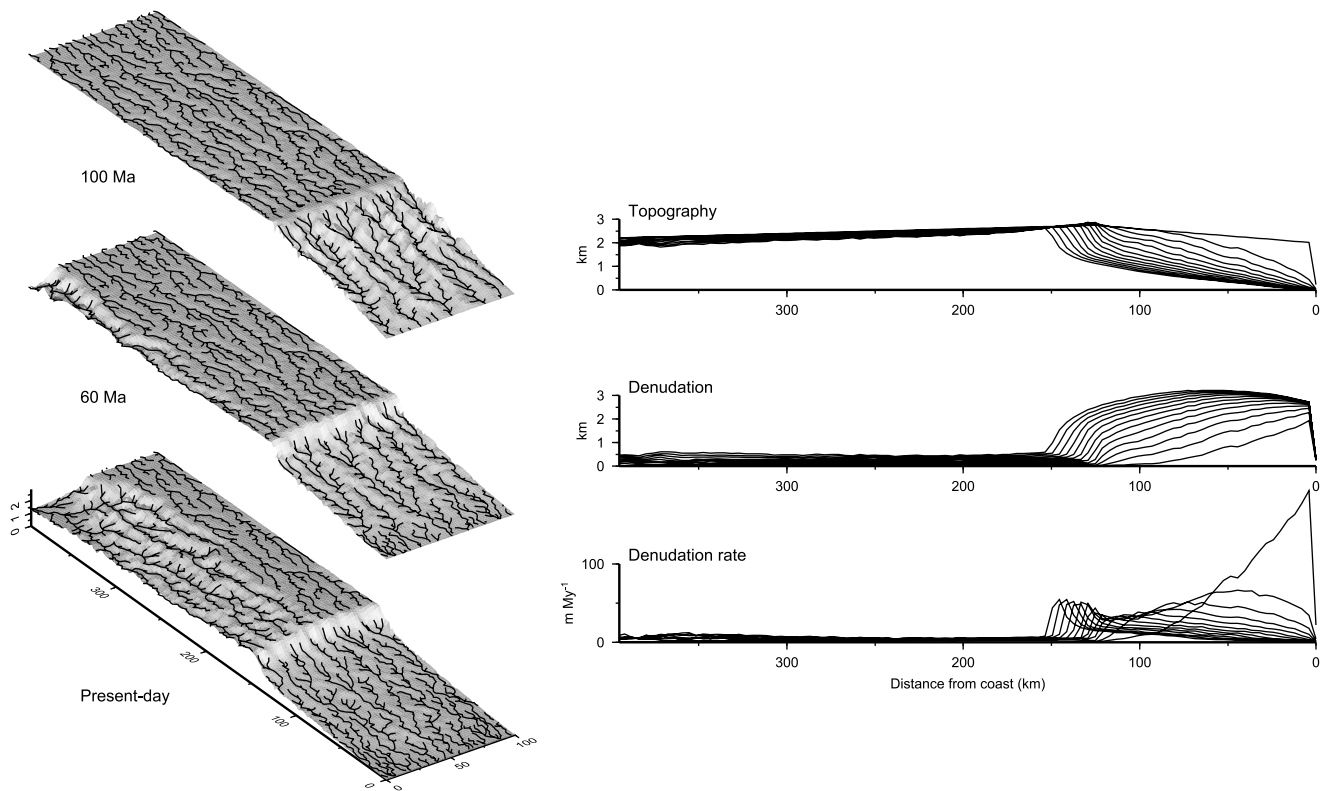


Figure 11. Predicted evolution for the inland base level fall model (400-km-long model in which 1 km lowering of the (0, 400 km) corner occurs between 85 and 80 Ma, other model parameters as in the lithological variation model). Artificially illuminated oblique views of the topography and drainage patterns at 100 Ma, 60 Ma, and the present-day (left) and the evolution of strike-averaged topography, denudation, and isostatic rebound at 10 m.y. intervals (right) are shown. See color version of this figure at back of this issue.

present-day escarpment, any relief that was created by rifting and breakup would be rapidly destroyed by rivers draining seaward, and replaced by an escarpment formed at the locus of the preexisting drainage divide. Once established, this newly formed escarpment will retreat at a rate determined in the model by the fluvial incision parameters K_{fVR} , and L_{fb} . Our preferred parameter sets are a compromise between fitting the long-term denudation rates inferred from the fission track data and the short-term escarpment retreat rates from the cosmogenic isotope data. A low value of the flexural rigidity will act to further slow down escarpment retreat rates by increasing the gradients of inland flowing streams and thus making them more efficient in competing with the seaward drainage.

[50] *Partridge and Maud* [1987] suggested that the escarpment evolved by parallel retreat, but that retreat rates diminished by an order of magnitude between the Early Cretaceous and the beginning of the Cenozoic. Whereas such a scenario may be compatible with the AFT data, the basis on which *Partridge and Maud* [1987] infer this significant decline in retreat rates does not appear to be robust. Moreover, they do not provide a physical mechanism for such a marked decline in retreat rates. In our models, escarpment retreat occurs mainly through fluvial incision and is thus dependent on both slope and drainage area. The destruction of the initial escarpment in the plateau degradation model is caused by rivers with large contribu-

ting drainage areas incising it in rapidly expanding gorges. Retreat rates of the gorge-heads decrease dramatically once they have reached the drainage divide and become pinned. In the escarpment retreat model, however, the escarpment always coincides with the drainage divide and therefore the contributing drainage areas of rivers incising the escarpment does not evolve over time. Escarpment retreat rates therefore remain constant (Figure 5). Moreover, a field study by *Seidl et al.* [1996] on the east Australian margin showed a lack of correlation between escarpment retreat rates and contributing areas of rivers, suggesting that gravitational slope processes (which are not dependent on area) rather than fluvial incision control escarpment retreat [*Weissel and Seidl*, 1998]. Climatic change has been claimed to be responsible for slowing down escarpment retreat [*Partridge*, 1997] but, to our present understanding, it is not clear how a change from a tropical Cretaceous to a humid subtropical Cenozoic climate would have slowed down the processes leading to escarpment retreat by an order of magnitude.

6.2. Cenozoic Uplift or Stability?

[51] Our inference that the southeast African margin has remained relatively stable since breakup is counter to the classic models of landscape evolution for this region [*King*, 1962; *Partridge and Maud*, 1987], which suggest that pulses of significant surface uplift have affected the margin during Cenozoic times.

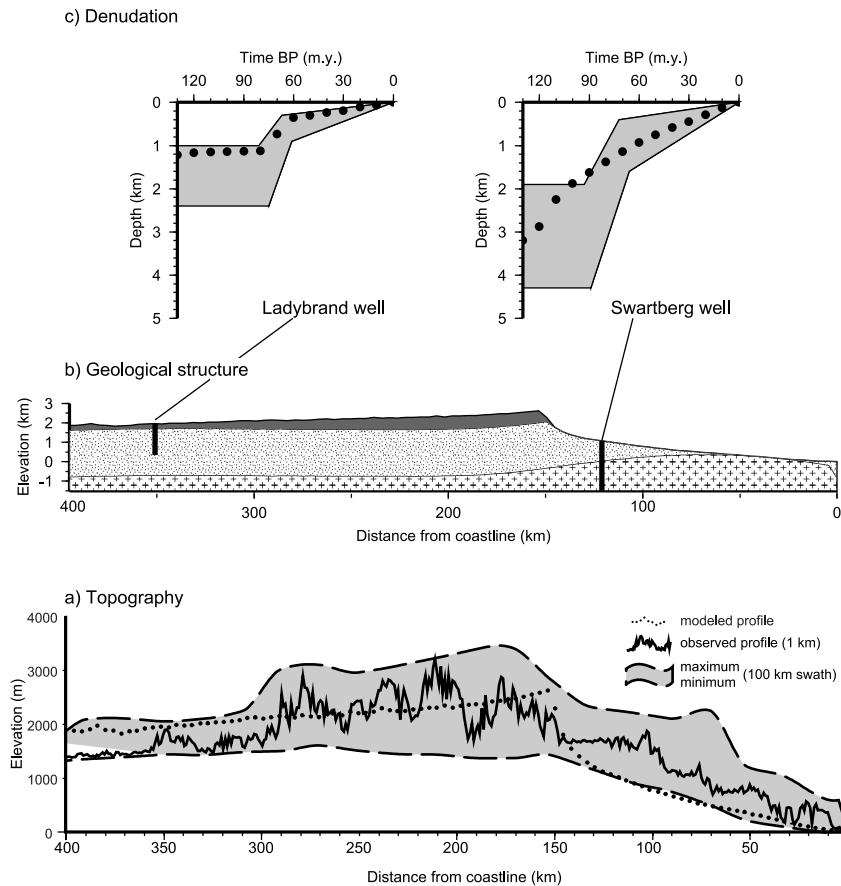


Figure 12. (a) Predicted present-day topography, compared to that observed and (b) predicted geological structure for the inland base level fall model. Shading in Figure 12b as in Figure 9. (c) Strike-averaged denudation history 30 km seaward (right) and 200 km inland (left) of the escarpment predicted by the model. Acceptable denudation histories from the Swartberg and Ladybrand well fission track data [cf. *Brown et al.*, 2002] are shown for comparison.

[52] The hypothesis of Cenozoic surface uplift is based on the correlation and dating of mapped surface remnants. However, the basis for correlating surface remnants and for prescribing ages to them has been questioned repeatedly [e.g., *Summerfield*, 1985, 1996; *Brown et al.*, 2000] (see section 3.2). Specifically, the step-like morphology of large parts of the southeast African margin may be lithologically controlled [e.g., *Wellington*, 1955]. Differences in erosional resistance of individual units within the upper Karoo sedimentary sequence appear to control the existence of subsidiary escarpments and ridges seaward of the main escarpment face. All of our models in which landform development occurs through the degradation of an initial plateau predict the existence of subsidiary ridges, either as long finger-like buttresses in the uniform lithology models or as narrower but well expressed flat-topped ridges in the models that include a resistant layer within the Karoo sequence.

[53] In subsequent publications, *Partridge et al.* [1995] and *Partridge* [1997] provided additional arguments for their inferred history of surface uplift and landscape development. These comprise (1) the existence of Cenozoic marine deposits at elevations up to a few hundred meters in western, southern, and southeastern South Africa; (2) the sedimentary record of the offshore Natal Basin [*Martin*, 1987]; and (3) fluvial long profiles of east coast rivers,

which show conspicuous knickpoints. None of these arguments, however, conclusively argues for Cenozoic uplift.

[54] Late Cretaceous and Eocene marine deposits occur at elevations of up to 400 m in the area between Port Elizabeth and East London, 400 km to the southwest of our study area. The elevations of many of these deposits are in fact lower than what our models would predict from the isostatic response to margin denudation combined with a reasonable estimate for the post-Cretaceous sea level fall (see Figure 8). This would suggest that the region where these deposits are preserved has in fact subsided with respect to the adjacent parts of the margin.

[55] Model predictions of isostatic rebound are strongly influenced by the choice of a specific elastic plate model. Our models are somewhat limited in that they employ a uniform-thickness continuous elastic plate, whereas T_e across the SE African margin probably varies strongly laterally, from >70 km on the Kaapvaal Craton to <10 km offshore [*Doucouré et al.*, 1996; *Hartley et al.*, 1996; *Watts and Marr*, 1995]. Incorporating such a laterally varying flexural rigidity in the models would lower the amount of isostatic rebound (and thus the predicted topography) inland but would not influence our predictions for the coastal part of the model. Likewise, incorporating a broken instead of a continuous elastic plate would not strongly influence our

modeling results, but would allow to reach the same amounts of isostatic rebound for somewhat higher flexural rigidities.

[56] Late Cenozoic marine sediments have been described from the southern coast where they occur at elevations of up to 300–400 m [Maud and Botha, 2000, and references therein]. These sediments are restricted to the Algoa Bay region and indicate that this part of the margin may have experienced local uplift exceeding the amount of flexural isostatic rebound we infer. However, the occurrence of correlative deposits at ≤ 50 m above sea level along the KwaZulu coast north of Durban [Maud and Botha, 2000] suggests that the east coast of southern Africa has been relatively stable during late Cenozoic times.

[57] Sediment accumulation rates in the Natal Basin for the past 5 m.y. are $3.96 \times 10^6 \text{ m}^3 \text{ yr}^{-1}$, only 6% higher than the long-term (postbreakup) rate of $3.73 \times 10^6 \text{ m}^3 \text{ yr}^{-1}$ [Martin, 1987, Table 2]. The 50% increase in sedimentation rate that Partridge *et al.* [1995] infer from Martin's [1987] study appears to be related to a large decrease in depocenter size because of cessation of sedimentation on the Mozambique coastal plain, rather than an increase in sediment flux from the onshore margin.

[58] Major knickpoints occurring in seaward flowing rivers of southeast Africa may be lithologically controlled, in the same way as the subsidiary flat-topped ridges. Our lithological variation model in fact predicts the development of knickpoints in major seaward flowing rivers, where they cross more resistant layers within the sedimentary sequence (Figure 10) which serve as local (secondary) base levels.

[59] The observations thus appear difficult to reconcile with scenarios involving several hundreds of meters of Cenozoic surface uplift and our modeling results show that such uplift is not required to explain the present-day morphology of the southeast African margin. Our inference that southeast Africa has been at a high elevation since at least the Early Cretaceous prompts the question of the nature of the uplift mechanism. Magmatic underplating associated with Early Jurassic Karoo volcanism has long been envisaged to have resulted in surface uplift of southern Africa and other intraplate regions [Cox, 1993; McKenzie, 1984]. Simple isostatic calculations, however, indicate that a very thick underplate (16.5–27.5 km for underplate densities of 2800–3000 kg m^{-3} and a mantle density of 3300 kg m^{-3}) is required to explain the 2.5 km initial elevation of the margin on its own. Recent seismological data [Nguiri *et al.*, 2001] do not provide evidence for such widespread and massive magmatic underplating beneath southern Africa. A more promising uplift mechanism, related to Karoo flood volcanism, would be the presence of large amounts of low-density melt residue in the lower mantle beneath southern Africa (see, for instance, discussion and references of Lowry *et al.* [2000]). Such low-density material would be characterized by high P wave velocities and contribute to the thick “tectosphere” beneath southern Africa [James *et al.*, 2001]; its relatively high viscosity would imply that it could remain stable for long time periods.

[60] Alternatively, Pysklywec and Mitrovica [1999] have suggested that dynamic rebound after slab detachment and cessation of subduction below the Cape Fold Belt active margin may have led to significant surface uplift at the end of the Triassic. If the entire 6-km-thick Karoo sedimentary sequence was laid down in a dynamically subsiding, sub-

duction-induced basin, then cessation of subduction-related subsidence would lead to 1.5–2.0 km of isostatic uplift (for a sediment density of 2400 kg m^{-3}), leaving <1 km of the initial elevation to be explained by other mechanisms. A similar mechanism has been proposed to explain uplift of the East Australian margin [Gallagher *et al.*, 1994a; Gurnis *et al.*, 1998]

[61] Finally, the anomalous topography of southern Africa may be dynamically supported by a large low-density body in the lower mantle (the “African superplume”) as suggested by Gurnis *et al.* [2000] and Lithgow-Bertelloni and Silver [1998], but this feature may be much longer-lived than previously assumed and the present-day dynamic uplift rates associated with it may be negligible (e.g., $\leq 1 \text{ m m.y.}^{-1}$).

[62] Our models satisfactorily reproduce the landscape development and denudation history of the SE African margin, although predicted rates and amounts of denudation are close to the minimum estimates of those recorded early in the history of the margin, during the Mid to Late Cretaceous (see Figure 12). A phase of rapid denudation during this period appears to be recorded throughout southern Africa and to coincide with a peak in offshore sedimentation rates [Brown *et al.*, 1994; Gallagher and Brown, 1999]. Our models do not predict a strong peak in denudation rates at this time but fairly constant rates of denudation throughout the first ~ 50 m.y. of the model runs. It has been argued that rift flank uplift is required to explain the total amounts of denudation recorded at passive margins [e.g., van der Beek *et al.*, 1994]. In the present case, however, rift flank uplift cannot be invoked because of the postrift timing and widespread occurrence of the apparent peak in denudation rates. Moreover, it is difficult to envisage how landscape development through plateau degradation can be reconciled with rift flank uplift, as the latter would install a new drainage divide that coincides with the escarpment [van der Beek and Braun, 1999]. Finally, our models predict the total amounts of postbreakup denudation reasonably well, but not the peak in denudation rates 40–70 m.y. after breakup. Whether this peak is real and, if so, what mechanism could explain it, is not clear at present.

[63] Although we will not pursue it in detail here, a remaining issue is the contrast between the present topography of the Lesotho Highlands, which have a high local relief, and the low rates of long-term denudation implied by the thermochronological data. Only a few km inland of the locally flat topography of the Drakensberg Escarpment rim, the headwaters of the Orange River have cut deep gorges creating a mean local relief of ~ 1 to 2 km [Summerfield, 1991]. This apparent discrepancy between long-term denudation rates and the present topography suggests that the current deeply incised landscape may have been created relatively recently through headward incision of the upper tributaries of the Orange River system cutting into the resistant Drakensberg Basalt. Further light may be shed on this problem using cosmogenic isotope analysis to quantify rates of river incision and valley-side slope retreat.

7. Conclusions

[64] Our numerical modeling results support a conceptual model for the postbreakup geomorphic development of the southeast African margin, in which a preexisting high-ele-

vation plateau experiences rapid erosion by rivers flowing seaward from a drainage divide located some 20–30 km to the east of the present-day escarpment. In such a model, a new escarpment forms at the locus of the drainage divide several tens of millions of years after breakup, and subsequently slowly retreats. Two key elements of this model contradict previous hypotheses for the evolution of the southeast African margin, based on the interpretation of erosion surfaces: (1) the escarpment has experienced only limited retreat since breakup, and was therefore initiated close to its present-day position rather than at the shoreline or the shelf edge; and (2) no large-scale Cenozoic tectonic surface uplift is required to explain the morphology and denudation history of the southeast African margin.

[65] Our models shed light on the major factors controlling postbreakup landscape evolution on passive margins. These factors have been explored before in a generic sense, but here their applicability to a specific setting, the southeast African margin, has been tested. The prebreakup topography of the margin appears to exert a fundamental control; models which include a preexisting drainage divide evolve by rapid degradation of the plateau surface seaward of the divide, whereas models without a preexisting drainage divide evolve by parallel escarpment retreat. Secondary controls are exerted by the flexural rigidity of the lithosphere, lithological variation in the eroded upper crustal section, and inland base level falls. Our modeling results suggest that for the southeast African margin, a relatively low effective elastic thickness of the lithosphere (T_e) of ~10 km explains the observed pattern of denudation as well as the observed geological structure. Lithological variations lead to the establishment of flat-topped ridges extending out from the escarpment as well as major knickpoints on rivers. Both these features have previously been interpreted as confirming Cenozoic surface uplift of the margin. An inland base level fall occurring 40–50 m.y. after breakup may explain the observed amounts of denudation inland of the escarpment as well as the development of inland drainage parallel to the escarpment. These model results suggest that care should be taken when using scenarios for surface uplift that are based on classic geomorphic studies as constraints on geodynamic models.

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References

- Beaumont, C., H. Kooi, and S. Willett, Progress in coupled tectonic—Surface process models with applications to rifted margins and collisional orogens, in *Geomorphology and Global Tectonics*, edited by M. A. Summerfield, pp. 29–56, John Wiley, New York, 2000.
- Ben-Avraham, Z., C. J. H. Hartnady, and J. A. Malan, Early tectonic extension between the Agulhas Bank and the Falkland Plateau due to the rotation of the Lafonia microplate, *Earth Planet. Sci. Lett.*, **117**, 43–58, 1993.
- Bierman, P. R., and M. Caffee, Slow rates of rock surface erosion and sediment production across the Namib desert and escarpment, southern Africa, *Am. J. Sci.*, **301**, 326–358, 2001.
- Bohannon, R. G., C. W. Naeser, D. L. Schmidt, and R. A. Zimmerman, The timing of uplift, volcanism and rifting peripheral to the Red Sea: A case for passive rifting?, *J. Geophys. Res.*, **94**, 1683–1701, 1989.
- Braun, J., and M. Sambridge, Modelling landscape evolution on geological time scales: A new method based on irregular spatial discretization, *Basin Res.*, **9**, 27–52, 1997.
- Brown, R. W., D. J. Rust, M. A. Summerfield, A. J. W. Gleadow, and M. C. J. de Wit, An Early Cretaceous phase of accelerated erosion on the southwestern margin of Africa: Evidence from apatite fission track analysis and the offshore sedimentary record, *Nucl. Tracks Rad. Meas.*, **17**, 339–350, 1990.
- Brown, R. W., M. A. Summerfield, and A. J. W. Gleadow, Apatite fission-track analysis: Its potential for the estimation of denudation rates and implications for models of long-term landscape development, in *Process Models and Theoretical Geomorphology*, edited by M. J. Kirkby, pp. 24–53, John Wiley, New York, 1994.
- Brown, R. W., K. Gallagher, A. J. W. Gleadow, and M. A. Summerfield, Morphotectonic evolution of the South Atlantic margins of Africa and South America, in *Geomorphology and Global Tectonics*, edited by M. A. Summerfield, pp. 257–283, John Wiley, New York, 2000.
- Brown, R. W., M. A. Summerfield, and A. J. W. Gleadow, Denudational history along a transect across the eastern margin (Drakensberg Escarpment) of southern Africa derived from apatite fission-track thermochronology, *J. Geophys. Res.*, **107**, 10.1029/2001JB000745, in press, 2002.
- Burke, K., The African Plate, *S. Afr. J. Geol.*, **99**, 339–410, 1996.
- Cockburn, H. A. P., R. W. Brown, M. A. Summerfield, and M. A. Seidl, Quantifying passive margin denudation and landscape evolution using a combined fission-track thermochronology and cosmogenic isotope analysis approach, *Earth Planet. Sci. Lett.*, **179**, 429–435, 2000.
- Cox, K. G., Continental magmatic underplating, *Philos. Trans. R. Soc. London, Ser. A*, **342**, 155–166, 1993.
- De Swardt, A. M. J., and G. Bennet, Structural and physiographic development of Natal since the Late Jurassic, *Trans. Geol. Soc. S. Afr.*, **77**, 309–322, 1974.
- Dingle, R. V., W. G. Siesser, and A. R. Newton, *Mesozoic and Tertiary Geology of Southern Africa*, 375 pp., A. A. Balkema, Brookfield, Vt., 1983.
- Dixey, F., Some aspects of the geomorphology of central and southern Africa, *Trans. Geol. Soc. S. Afr.*, **58** (Annex), 1955.
- Doucouré, C. M., M. J. de Wit, and M. F. Mushayandebvu, Effective elastic thickness of the continental lithosphere in South Africa, *J. Geophys. Res.*, **101**, 11,291–11,303, 1996.
- Duncan, R. A., P. R. Hooper, J. Rehacek, J. S. Marsh, and A. R. Duncan, The timing and duration of the Karoo igneous event, southern Gondwana, *J. Geophys. Res.*, **102**, 18,127–18,138, 1997.
- Fleming, A., River long profiles: Implications for cyclical explanations of landscape evolution in South Africa, in *Proceedings of the 4th International Conference on Geomorphology, Bologna (Italy)* 28/08-03/09/1997, *Geogr. Fis. Din. Quat. Suppl.*, **III**, 167, 1997.
- Fleming, A., Landscape development in the southern Drakensberg and Lesotho Highlands, SE Africa: Quantifying denudation rates using in situ-produced cosmogenic chlorine-36 data, Ph.D. thesis, 323 pp. Univ. of Edinburgh, Edinburgh, U.K., 2000.
- Fleming, A., M. A. Summerfield, J. O. Stone, L. K. Fifield, and R. G. Cresswell, Denudation rate for the southern Drakensberg escarpment, SE Africa, derived from in-situ-produced cosmogenic ^{36}Cl : Initial results, *J. Geol. Soc. London*, **156**, 209–212, 1999.
- Gadd, S. A., and R. A. Scrutton, An integrated thermomechanical model for transform continental margin evolution, *Geo Mar. Lett.*, **17**, 21–30, 1997.
- Gallagher, K., and R. W. Brown, The onshore record of passive margin evolution, *J. Geol. Soc. London*, **154**, 451–457, 1997.
- Gallagher, K., and R. W. Brown, The Mesozoic denudation history of the Atlantic margins of southern Africa and southeast Brazil and the relationship to offshore sedimentation, in *The Oil and Gas Habitats of the South Atlantic*, edited by N. R. Cameron, R. H. Bate, and V. S. Clure, *Geol. Soc. Spec. Publ.*, **153**, 41–53, 1999.
- Gallagher, K., T. A. Dumitru, and A. J. W. Gleadow, Constraints on the vertical motion of eastern Australia during the Mesozoic, *Basin Res.*, **6**, 77–94, 1994a.
- Gallagher, K., C. J. Hawkesworth, and M. J. M. Mantovani, The denudation history of the onshore continental margin of SE Brazil inferred from apatite fission track data, *J. Geophys. Res.*, **99**, 18,117–18,145, 1994b.
- Gilchrist, A. R., and M. A. Summerfield, Differential denudation and flexural isostasy in formation of rifted-margin upwarps, *Nature*, **346**, 739–742, 1990.
- Gilchrist, A. R., and M. A. Summerfield, Denudation, isostasy and landscape evolution, *Earth Surf. Processes Landforms*, **16**, 555–562, 1991.

- Gilchrist, A. R., and M. A. Summerfield, Tectonic models of passive margin evolution and their implications for theories of long-term landscape development, in *Process Models and Theoretical Geomorphology*, edited by M. J. Kirkby, pp. 221–246, John Wiley, New York, 1994.
- Gilchrist, A. R., H. Kooi, and C. Beaumont, The post-Gondwana geomorphic evolution of southwestern Africa: Implications for the controls on landscape development from observations and numerical experiments, *J. Geophys. Res.*, **99**, 12,211–12,228, 1994.
- Gunnell, Y., The interaction between geological structure and global tectonics in multistoreyed landscape development: A denudation chronology of the South Indian shield, *Basin Res.*, **10**, 281–310, 1998.
- Gurnis, M., R. D. Mueller, and L. Moresi, Dynamics of Cretaceous vertical motion of Australia and the Australian-Antarctic discordance, *Science*, **279**, 1499–1504, 1998.
- Gurnis, M., J. X. Mitrovica, J. Ritsema, and H.-J. van Heijst, Constraining mantle density structure using geological evidence of surface uplift rates: The case of the African Superplume, *Geochim. Geophys. Geosyst.*, **1**(Article), 1999GC000035 [26,963], 2000.
- Haq, B. U., J. Hardenbol, and P. R. Vail, Chronology of fluctuating sea levels since the Triassic, *Science*, **235**, 1156–1166, 1987.
- Hartley, R., A. B. Watts, and J. D. Fairhead, Isostasy of Africa, *Earth Planet. Sci. Lett.*, **137**, 1–18, 1996.
- Heimsath, A. M., W. E. Dietrich, K. Nishiizumi, and R. C. Finkel, The soil production function and landscape equilibrium, *Nature*, **388**, 358–361, 1997.
- James, D. E., M. J. Fouch, and J. C. VanDecar, Tectospheric structure beneath southern Africa, *Geophys. Res. Lett.*, **28**, 2485–2488, 2001.
- King, L. C., *South African Scenery*, 699 pp., Oliver and Boyd, White Plains, N. Y., 1951.
- King, L. C., Pediplanation and isostasy: An example from South Africa, *Q. J. Geol. Soc. London*, **111**, 353–359, 1955.
- King, L. C., *The Morphology of the Earth*, 726 pp., Oliver and Boyd, White Plains, N. Y., 1962.
- King, L. C., *The Natal Monocline*, 134 pp., Univ. of Natal Press, Pietermaritzburg, South Africa, 1982.
- Kooi, H., and C. Beaumont, Escarpment evolution on high-elevation rifted margins: insights derived from a surface processes model that combines diffusion, advection and reaction, *J. Geophys. Res.*, **99**, 12,191–12,210, 1994.
- Kooi, H., and C. Beaumont, Large-scale geomorphology: Classical concepts reconciled and integrated with contemporary ideas via a surface-processes model, *J. Geophys. Res.*, **101**, 3361–3386, 1996.
- Lithgow-Bertelloni, C., and P. G. Silver, Dynamic topography, plate driving forces and the African superswell, *Nature*, **395**, 269–272, 1998.
- Lowry, A. R., N. M. Ribe, and R. B. Smith, Dynamic elevation of the Cordillera, western United States, *J. Geophys. Res.*, **105**, 23,371–23,390, 2000.
- Martin, A. K., Comparison of sedimentation rates in the Natal Valley, south-west Indian Ocean, with modern sediment yields in east coast rivers of southern Africa, *S. Afr. J. Sci.*, **83**, 716–724, 1987.
- Martin, A. K., and C. J. H. Hartnady, Plate tectonic development of the southwest Indian Ocean: A revised reconstruction of East Antarctica and Africa, *J. Geophys. Res.*, **91**, 4767–4786, 1986.
- Maud, R. R., A preliminary review of the structure of coastal Natal, *Trans. Geol. Soc. S. Afr.*, **64**, 247–256, 1961.
- Maud, R. R., and G. A. Botha, Deposits of the south eastern and southern coasts, in *The Cenozoic of Southern Africa*, edited by T. C. Partridge and R. R. Maud, pp. 19–32, Oxford Univ. Press, New York, 2000.
- McKenzie, D. P., A possible mechanism of epeirogenic uplift, *Nature*, **307**, 616–619, 1984.
- Menzies, M., K. Gallagher, A. Yelland, and A. J. Hurford, Volcanic and nonvolcanic rifted margins of the Red Sea and Gulf of Aden: Crustal cooling and margin evolution in Yemen, *Geochim. Cosmochim. Acta*, **61**, 2511–2527, 1997.
- Nguiri, T. K., J. Gore, D. E. James, S. J. Webb, C. Wright, T. G. Zengeni, O. Gwavava, and J. A. Snoke, Crustal structure beneath southern Africa and its implications for the formation and evolution of the Kaapvaal and Zimbabwe cratons, *Geophys. Res. Lett.*, **28**, 2501–2504, 2001.
- Nyblade, A. A., and S. W. Robinson, The African Superswell, *Geophys. Res. Lett.*, **21**, 765–768, 1994.
- Ollier, C. D., (Ed.), Morphotectonics of passive continental margins, *Z. Geomorph. Suppl.*, **54**, 120 pp., 1985.
- Ollier, C. D., and M. E. Marker, The Great Escarpment of southern Africa, *Z. Geomorph. Suppl.*, **54**, 37–56, 1985.
- Ollier, C. D., and C. F. Pain, Equating the basal unconformity with the palaeoplain: A model for passive margins, *Geomorphology*, **19**, 1–15, 1997.
- Oreskes, N., K. Shrader-Frechette, and K. Belitz, Verification, validation, and confirmation of numerical models in the earth sciences, *Science*, **263**, 641–646, 1994.
- Partridge, T. C., Late Neogene uplift in eastern and southern Africa and its paleoclimatic implications, in *Tectonic Uplift and Climate Change*, edited by W. F. Ruddiman, pp. 63–85, Plenum, New York, 1997.
- Partridge, T. C., and R. R. Maud, Geomorphic evolution of southern Africa since the Mesozoic, *S. Afr. J. Geol.*, **90**, 179–208, 1987.
- Partridge, T. C., and R. R. Maud, Macroscale geomorphic evolution of southern Africa, in *The Cenozoic of Southern Africa*, edited by T. C. Partridge and R. R. Maud, pp. 3–18, Oxford Univ. Press, New York, 2000.
- Partridge, T. C., B. A. Wood, and P. B. deMenocal, The influence of global climatic change and regional uplift on large-mammalian evolution in east and southern Africa, in *Paleoclimate and Evolution, with Emphasis on Human Origins*, edited by E. S. Vrba et al., pp. 331–355, Yale Univ. Press, New Haven, Conn., 1995.
- Pitman, W. C., III, The relationship between eustasy and stratigraphic sequences of passive margins, *Geol. Soc. Am. Bull.*, **89**, 1389–1403, 1978.
- Pugh, J. C., Isostatic readjustment in the theory of pediplanation, *Q. J. Geol. Soc. London*, **111**, 361–369, 1955.
- Pysklyvec, R. N., and J. X. Mitrovica, The role of subduction-induced subsidence in the evolution of the Karoo Basin, *J. Geol.*, **107**, 155–164, 1999.
- Rust, D. J., and M. A. Summerfield, Isopach and borehole data as indicators of rifted margin evolution in southwestern Africa, *Mar. Pet. Geol.*, **7**, 277–287, 1990.
- Seidl, M. A., J. K. Weisell, and L. F. Pratson, The kinematics and pattern of escarpment retreat across the rifted continental margin of SE Australia, *Basin Res.*, **8**, 301–316, 1996.
- Smith, R. H. M., A review of stratigraphy and sedimentary environments of the Karoo Basin of South Africa, *J. Afr. Earth Sci.*, **10**, 117–137, 1990.
- Stock, J. D., and D. R. Montgomery, Geologic constraints on bedrock river incision using the stream power law, *J. Geophys. Res.*, **104**, 4983–4993, 1999.
- Stüwe, K., Flexural constraints on the denudation of asymmetric mountain belts, *J. Geophys. Res.*, **96**, 10,401–10,408, 1991.
- Summerfield, M. A., Plate tectonics and landscape evolution on the African continent, in *Tectonic Geomorphology*, edited by M. Morisawa and J. T. Hack, pp. 27–51, Allen and Unwin, Concord, Mass., 1985.
- Summerfield, M. A., Sub-aerial denudation of passive margins: Regional elevation versus local relief models, *Earth Planet. Sci. Lett.*, **102**, 460–469, 1991.
- Summerfield, M. A., Tectonics, geology, and long-term landscape development, in *The Physical Geography of Africa*, edited by W. M. Adams, A. S. Goudie, and A. R. Orme, pp. 1–17, Oxford Univ. Press, New York, 1996.
- Summerfield, M. A., Geomorphology and global tectonics: Introduction, in *Geomorphology and Global Tectonics*, edited by M. A. Summerfield, pp. 3–11, John Wiley, New York, 2000.
- Summerfield, M. A., and R. W. Brown, Geomorphic factors in the interpretation of fission-track data, in *Advances in Fission-Track Geochronology*, edited by P. Van den Haute and F. De Corte, pp. 269–284, Kluwer Acad., Norwell, Mass., 1998.
- ten Brink, U., and T. Stern, Rift flank uplifts and hinterland basins: Comparison of the Transantarctic Mountains with the Great Escarpment of southern Africa, *J. Geophys. Res.*, **97**, 569–585, 1992.
- Tucker, G. E., and R. L. Slingerland, Erosional dynamics, flexural isostasy, and long-lived escarpments: A numerical modeling study, *J. Geophys. Res.*, **99**, 12,229–12,243, 1994.
- van Balen, R. T., P. A. van der Beek, and S. Cloetingh, The effect of rift shoulder erosion on stratigraphic patterns at passive margins: Implications for sequence stratigraphy, *Earth Planet. Sci. Lett.*, **134**, 527–544, 1995.
- van der Beek, P. A., and J. Braun, Numerical modelling of landscape evolution on geological time-scales: A parameter analysis and comparison with the south-eastern highlands of Australia, *Basin Res.*, **10**, 49–68, 1998.
- van der Beek, P. A., and J. Braun, Controls on post-mid-Cretaceous landscape evolution in the southeastern highlands of Australia: Insights from numerical surface process models, *J. Geophys. Res.*, **104**, 4945–4966, 1999.
- van der Beek, P. A., S. Cloetingh, and P. A. M. Andriessen, Mechanisms of extensional basin formation and vertical motions at rift flanks: Constraints from tectonic modelling and fission-track thermochronology, *Earth Planet. Sci. Lett.*, **121**, 417–433, 1994.
- van der Beek, P. A., P. A. M. Andriessen, and S. Cloetingh, Morphotectonic evolution of rifted continental margins: Inferences from a coupled tectonic-surface processes model and fission-track thermochronology, *Tectonics*, **14**, 406–421, 1995.
- van der Wateren, F. M., and T. J. Dunai, Late Neogene passive margin denudation history—Cosmogenic isotope measurements from the central Namib desert, *Global Planet. Change*, **30**, 271–307, 2001.
- Vasconcelos, P. M., K-Ar and ⁴⁰Ar/³⁹Ar geochronology of weathering processes, *Annu. Rev. Earth Planet. Sci.*, **27**, 183–229, 1999.

- von Veh, M. W., and N. J. B. Anderson, Normal-slip faulting in the coastal areas of northern Natal and Zululand, South Africa, *S. Afr. J. Geol.*, 93, 574–582, 1990.
- Watts, A. B., and C. Marr, Gravity anomalies and the thermal and mechanical structure of rifted continental margins, in *Rifted Ocean-Continent Boundaries*, edited by E. Banda, pp. 65–94, Kluwer Acad., Norwell, Mass., 1995.
- Weissel, J. K., and M. A. Seidl, Inland propagation of erosional escarpments and river profile evolution across the southeast Australian passive continental margin, in *Rivers over Rock: Fluvial Processes in Bedrock Channels*, *Geophys. Monogr. Ser.*, vol. 107, edited by K. J. Tinkler and E. E. Wohl, pp. 189–206, Washington, D. C., 1998.
- Wellington, J. H., *Southern Africa: A Geographical Study*, vol. I, *Physical Geography*, 528 pp., Cambridge Univ. Press, New York, 1955.
- Widdowson, M., Tertiary palaeosurfaces of the SW Deccan, western India: Implications for passive margin uplift, in *Palaeosurfaces: Recognition, Reconstruction and Palaeoenvironmental Interpretation*, edited by M. Widdowson, *Geol. Soc. Spec. Publ.*, 120, 221–248, 1997.
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- J. Braun, Research School of Earth Sciences, Australian National University, Canberra ACT 0200, Australia. (Jean.Braun@anu.edu.au)
- R. W. Brown, School of Earth Sciences, University of Melbourne, Melbourne, VIC 3010, Australia.
- A. Fleming and M. A. Summerfield (corresponding author), Department of Geography, University of Edinburgh, Drummond Street, Edinburgh, EH8 9XP, UK. (mas@geo.ed.ac.uk)
- P. van der Beek, Laboratoire de Géodynamique des Chaînes Alpines, Université Joseph Fourier, BP 53, F-38041 Grenoble Cedex, France. (pvdbeek@ujf-grenoble.fr)

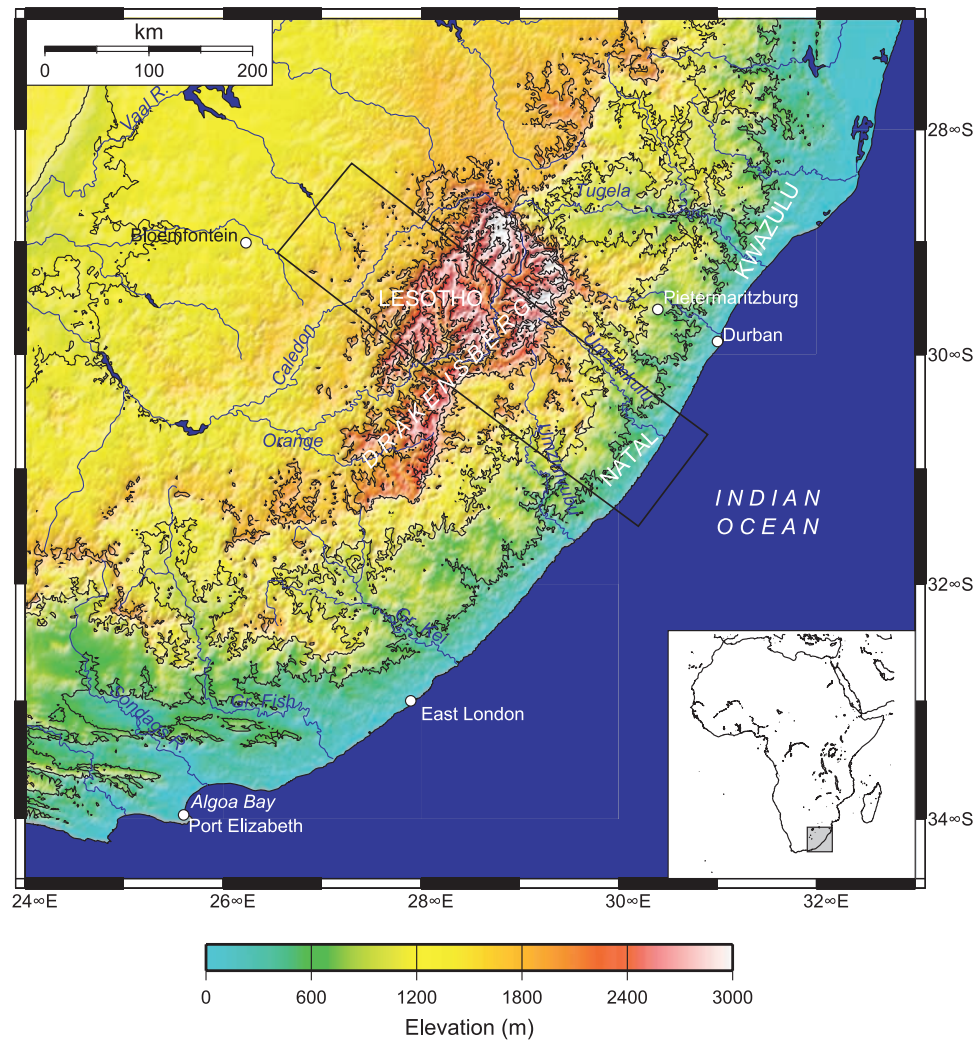


Figure 1. Shaded relief map of the southeastern African margin, color coded according to elevation and showing localities referred to in the text. Box indicates location of transect shown in Figure 2; inset map shows location of study area on the African continent.

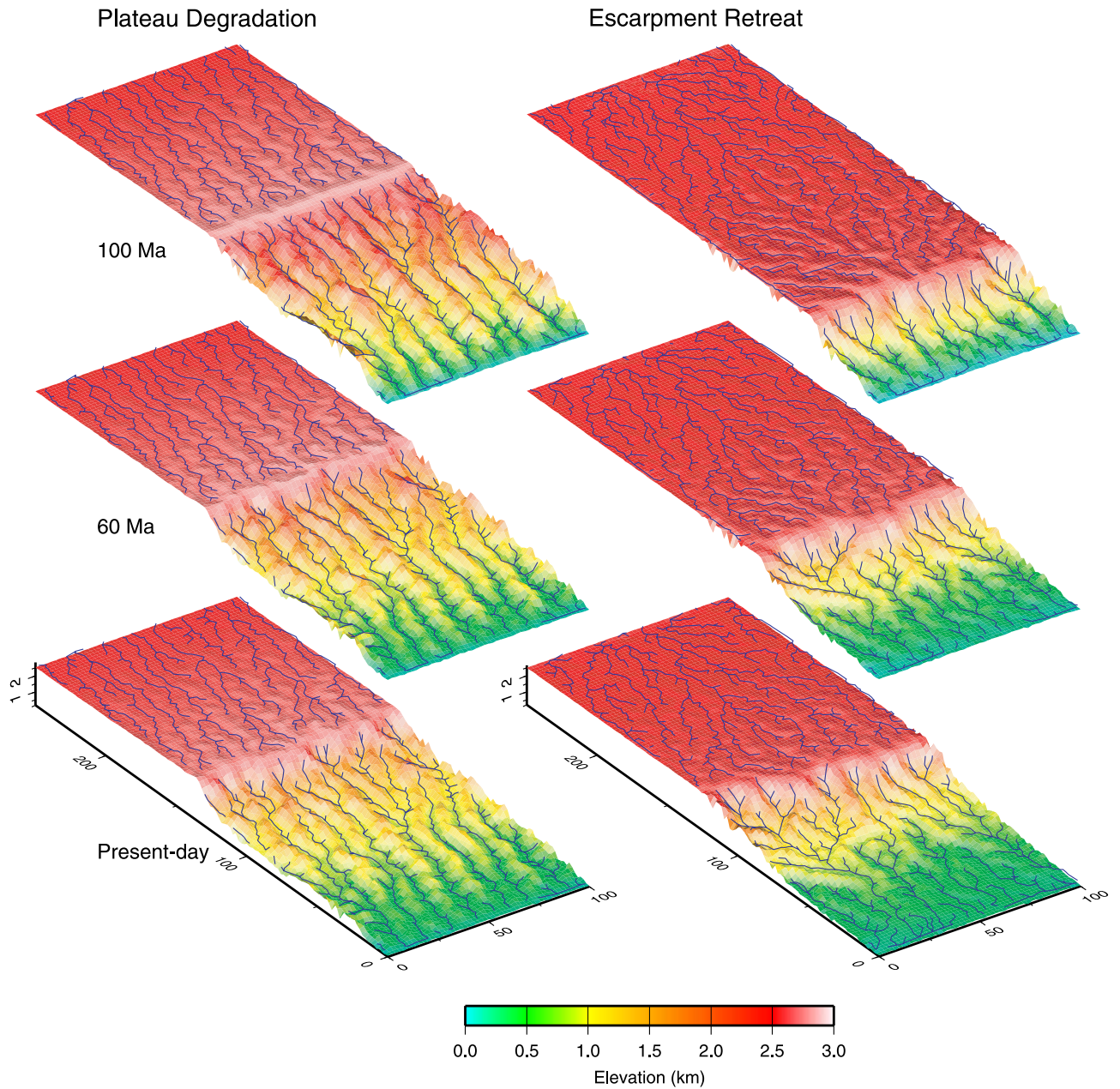


Figure 3. Artificially illuminated oblique views of the topography and drainage patterns predicted by the plateau degradation and escarpment retreat models, at 100 Ma, 60 Ma, and the present-day. Axes indicate model coordinates in kilometers. Parameter values for these models are given in Table 1.

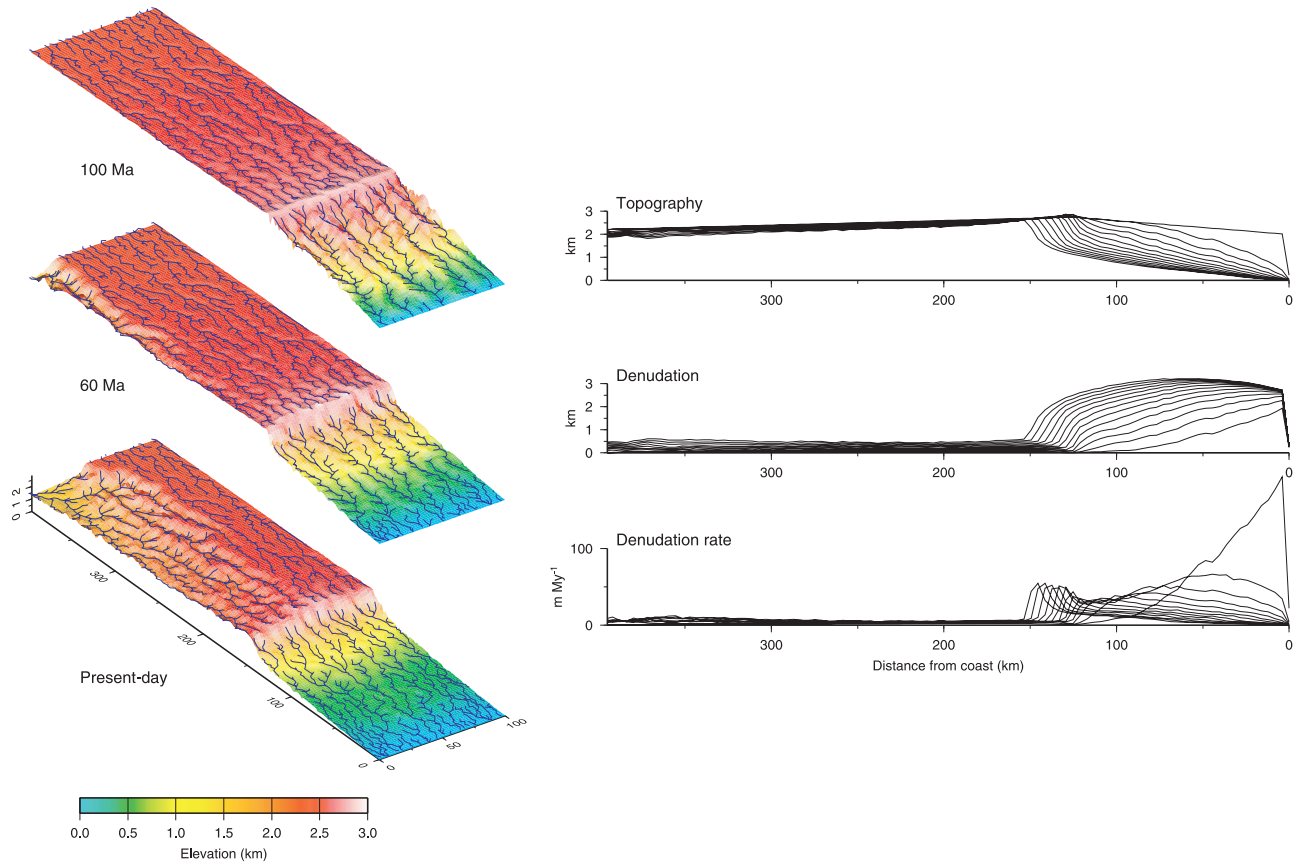


Figure 11. Predicted evolution for the inland base level fall model (400-km-long model in which 1 km lowering of the (0, 400 km) corner occurs between 85 and 80 Ma, other model parameters as in the lithological variation model). Artificially illuminated oblique views of the topography and drainage patterns at 100 Ma, 60 Ma, and the present-day (left) and the evolution of strike-averaged topography, denudation, and isostatic rebound at 10 m.y. intervals (right) are shown.